Kh. P. Pogosyan

THE AIR ENVELOPE OF THE EARTH

TRANSLATED FROM RUSSIAN

Published for the U.S. National Aeronautics and Space Administration
and the National Science Foundation, Washington, D.C.
by the Israel Program for Scientific Translations

Kh. P. POGOSYAN

THE AIR ENVELOPE OF THE EARTH

(Vozdushnaya obolochka zemli)

GIMIZ
Gidrometeorologicheskoe Izdatel'stvo
Leningrad 1962

Translated from Russian

Israel Program for Scientific Translations
Jerusalem 1965

NASA TT-F-287
TT 65-50113

Published Pursuant to an Agreement with
THE NATIONAL AERONAUTICS AND SPACE ADMINISTRATION, U. S. A.
and
THE NATIONAL SCIENCE FOUNDATION, WASHINGTON, D.C.

IPST Cat. No. 1397

Translated by I. Shechtman

Printed in Jerusalem by S. Monson
Binding: K. Wiener

Price: $ 6.00

Available from the
U.S. DEPARTMENT OF COMMERCE
Clearinghouse for Federal Scientific and Technical Information
Springfield, Va. 22151

IX/18/3.5

TABLE OF CONTENTS

ANNOTATION

Our knowledge of the atmosphere has broadened considerably in recent years. This book gives new information on the structure and composition of the atmosphere, the inflow of solar energy and the formation of temperature at various heights, and the motion of the air — the circulation of the atmosphere. Some chapters deal with such interesting phenomena as atmospheric fronts, jet streams, and powerful vortexes — cyclones and anticyclones — giving rise to sharp weather variations. The reader is introduced to modern weather forecasting and to the various possibilities of artificially influencing atmospheric processes in order to modify the weather.

There is a special chapter dealing with moisture circulation in the atmosphere, moisture rotation on land, and the distribution of precipitation over the globe.

A number of chapters contain data on the temperature at the surface and above the Earth that were obtained from observations made during the last three to five years, as well as data on changes in climate.

The book is intended for a wide circle of readers.

A word from the author

The technological progress which marks our times serves the natural sciences. Meteorology, the science of physical processes and phenomena observed in the atmosphere, is advancing rapidly.

During the 1940's and 50's meteorology was enriched by new means of observation, mainly radiotechnical. Since the second world war, particularly during the period of the International Geophysical Year (IGY), new data concerning the structure of the atmosphere and processes taking place at great heights have been revealed. Thousands of scientists from 60 countries took part in the extensive program.

Few areas now remain that are not covered by meteorological observations. This was not the case a short time ago. The atmospheric sounding network in the southern hemisphere has gradually expanded. Even on the once deserted icy continent of Antarctica, meteorological and aerological stations have been established. Here, 24-hour observations of the weather, temperature, pressure, humidity, and wind to heights of up to 30–40 km are carried out. New information on the upper layers of the atmosphere has been obtained by means of rockets and artificial satellites.

The new data opened a wide field for investigation of atmospheric processes with the aim of solving many practical problems in meteorology. Among these is the problem of weather forecasting, particularly long-range forecasting, as well as that of reducing the disastrous effects of some weather phenomena by artificially influencing them, etc.

The author has tried to outline the new achievements in the study of the lower and upper layers of the atmosphere – the troposphere and the stratosphere.

The book is a revised and completed edition of "Ocherkov ob atmosfere" (Essays on the Atmosphere) published in 1955. The author will consider his task fulfilled to some extent if he is able to satisfy the interest of the reader in the science of the atmosphere.

INTRODUCTION

Solar energy is the source of life on Earth. Without the atmosphere, however, our planet would have been as lifeless as its satellite the Moon. Having let through the solar radiation, the atmosphere then captures that part which is radiated by and reflected from the Earth, thus protecting the Earth from cooling. Clouds form in the atmosphere, precipitate, and complex processes develop which determine the character of the weather and climate.

The weather and climate exert a direct influence on our life and activities which take place on the bottom of the air "ocean." The ever-increasing interest in the various atmospheric phenomena and the wish not only to master the laws controlling them, but also to learn to predict the weather are thus understandable.

To study and understand weather phenomena and the processes developing in the atmosphere required prolonged and intensive effort. Unable to explain the origin of various and, in particular, dangerous, weather phenomena, people for many centuries read into them the manifestation of the power and will of supernatural forces. Gradually mastering the secrets of nature, man began to interpret weather phenomena correctly.

The history of meteorology is more than 2000 years old. The first book on meteorology was written in the fourth century B.C. by Aristotle. However, in spite of the fact that meteorology began so early in man's history, it developed only in the last two to three centuries. This development was aided by the invention of instruments for measuring the pressure, temperature, and other parameters that characterize the state of the atmosphere, as well as by the establishment and development of a network of meteorological stations; today they exceed ten thousand.

An equally important role in the development of meteorology was played by the new means of observation, which made it possible to penetrate the higher layers of the atmosphere. Sounding the atmosphere, and primarily radiosounding, gave many possibilities for obtaining information on the structure and composition of the atmosphere, on air currents, and other atmospheric characteristics. For example, it was established that in the upper troposphere and in the stratosphere strong air currents (the so-called jet streams) with velocities frequently reaching 300 km/hr and more, exist and develop continuously.

The study of the higher layers of the atmosphere, in particular of the stratosphere, has more than a pure scientific purpose. Airplanes, which depend on the direction and velocity of the wind, fly in the stratosphere. Wind and temperature data are also used for exact calculations of rocket flight.

With the development of meteorology in the 20th century, separate subsections appeared, i.e., climatology, aerology, synoptic and dynamic meteorology, actinometry, atmospheric optics, and atmospheric electricity.

The basic scientific investigations in meteorology are directed mainly at developing new and improving existing methods for weather prediction. However, the study of atmospheric processes for practical purposes includes more than just developing methods for weather forecasting.

In the Soviet Union, vast plans for transforming nature are being implemented. Gigantic dams and hydroelectric stations, water reservoirs, and irrigation channels are being constructed. Forest belts are being planted to protect fields, arid steppes and deserts are being supplied with water and irrigated. By investigating the peculiarities of the meteorological and hydrological regime using experimental and theoretical geophysics and its allied sciences, it is possible to give practical recommendations for raising the efficiency of climate melioration.

The study of the mechanism of atmospheric processes and of the regularities of their development makes it possible to reduce the harmful effects of various weather phenomena by direct action on the processes causing them.

The extensive investigations of the atmosphere in the USSR are based on the data of various types of observations; these are carried out not only in inhabited regions, but also in the taiga, in deserts, and on mountains. For a number of years, Soviet Arctic and Antarctic expeditions, with the most modern equipment, have been studying the meteorological and hydrological processes over huge territories of the Arctic and the Antarctic.

New information on the processes developing in the troposphere and in the stratosphere made it necessary to reconsider a number of theoretical assumptions on the general circulation of the atmosphere. It was found that the air exchange between the Equator and the Poles is not by circulation "wheels" between adjacent latitude belts of the Earth (Equator-tropics, tropics-middle latitudes, middle-high latitudes), but mainly by horizontal transfer of the air masses. In particular, the idea of a closed ring of trade wind circulation was proved wrong.

The idea of the stratosphere as a relatively quiet isothermal medium has also changed. New data showed that the temperature in the stratosphere undergoes considerable variations, which are particularly large at high latitudes, from season to season and even during each season.

However, there still remain many unsolved problems in meteorology. Among them the important Sun-atmosphere problem.

It is known that the atmosphere of the Earth reacts to processes taking place on the Sun. For example, the appearance of the aurora polaris at high latitudes, the disturbance of the geomagnetic field causing a disruption of radiocommunications, etc., are caused by processes on the Sun. Since the main source for the appearance and existence of atmospheric circulation is the solar energy, the mechanism of its transformation into kinetic energy in the atmosphere must be known. It is obvious that the processes on the Sun exert an influence not only on the general circulation of the atmosphere, but also on the weather. However, until now there has been no satisfactory theory as to the nature of this interaction. At the same time, understanding the interaction between processes developing in various layers of the atmosphere is particularly necessary for developing and improving methods for weather forecasting. These, as well as other important problems of meteorology, still need to be solved.

During the 40's and 50's of the 20th century, the amount of meteorological and aerological experimental data increased continuously. Whereas in the 30's almost the only basis for preparing the weather forecasting for a day ahead was the synoptic surface weather map, now, in addition to a number of surface maps, the forecaster always has at his disposal a large number of high altitude maps and numerous auxiliary maps, parts of which are obtained theoretically.

With the increase in data, the more difficult it became to process and use them in time for the preparation of the weather forecasting. Hydrodynamics came to help the forecaster.

At the beginning of the 40's experiments began, for the first time in the Soviet Weather Service, for predicting meteorological elements (temperature and pressure). Since then, hydrodynamic methods developed considerably and are now incorporated into operational prognostication and investigation work. Numerical methods are used in the preparation both of daily weather forecasts and of long-range forecasts for several days, a month, or even a season, using high-speed electronic computers that are capable of solving rapidly a large number of complicated equations with many unknowns.

Methods of weather prediction are promising, although at present they are only at the initial stage of their development.

The computer techniques are gradually finding application in climatology. The high-speed processing of observation data makes interesting generalizations possible and new regularities of climate formation available. By mechanizing the processing of climatological data, long-range (ten-day, month, season) statistical weather forecasting methods have been developed.

Modern meteorology has expanded and already possesses the basis for formulating and solving the most important problems of weather and climate, which play an increasingly important role in man's economic acitvity.

THE EARTH'S AIR ENVELOPE

The atmosphere

While rotating about its own axis, the Earth rushes around the Sun along an elliptic orbit with a mean velocity of 107,280 km/hr. The polar radius of the Earth is 6357 km, the equatorial radius, 6378 km, and the area of the Earth's surface, 510,100,900 km^2.

The Earth's atmosphere extends to heights of about 2000 km, i.e., its height amounts to about 1/3 of the radius of the Earth. However, the atmosphere does not have a clear boundary, its upper layers being highly rarefied.

The whole atmosphere weighs 5.3 quadrillions (5,300,000,000,000,000 tons). To get an idea of this amount, the atmosphere is 13 times heavier than the Caucasus Range and one million times lighter than the Earth.

The density of the atmosphere varies at different heights. The denser lower layers of the air are considerably heavier than the upper rarefied ones. The mass of the lower 5-kilometer layer is half the mass of the entire atmosphere. As one goes higher, the density as well as the composition, temperature, wind direction and velocity, and other atmospheric parameters vary considerably.

In accordance with the temperature variation with height, the following division of the atmosphere into layers is customary:

Layer	Height of the lower and upper boundaries	Transitional layer
Troposphere	From the Earth's surface to 8–17 km	Tropopause
Stratosphere	From the tropopause to 50–60 km	Stratopause
Mesosphere	From 50–60 km to 80 km	Mesopause
Thermosphere	From 80 to 700–800 km	Thermopause
Exosphere	Above 800 km	—

In the troposphere the temperature drops with increasing height. In the stratosphere, isothermal conditions or temperature rise with height prevails. A distinctive feature of the mesosphere is the temperature drop with increasing height. In the thermosphere the temperature rises almost continuously with height. The last, external sphere (exosphere) extends up to the boundary of the atmosphere.

We now give a more detailed description of the distinctive features of these layers, particularly of the troposphere and stratosphere, which are better known and understood than the higher-lying layers.

Composition of the atmosphere

The atmosphere consists of a mechanical mixture of various gases. If we remove the moisture and dust particles from the air, the dry air near the surface of the Earth will have the following volumetric composition: 78.09% nitrogen, 20.95% oxygen, 0.93% argon, 0.03% carbon dioxide. The remaining gases, i.e., hydrogen, neon, helium, krypton, xenon, ammonium, hydrogen peroxide, iodine, radium emanation, ozone, appear in negligible amounts.

The amount of carbon dioxide in the atmosphere varies. There is more in industrial regions than far from them. For example, the air above the Antarctic contains about 0.02% carbon dioxide, i.e., about 1/3 the average amount. In the ground layer the amount of carbon dioxide undergoes daily variations. At night it is somewhat larger than during the day because carbon dioxide is absorbed by plants only during the day, and at night the absorption stops. Industries and animals produce it continuously.

Measurements show that the burning of carbon during the present century has increased the total amount of carbon dioxide in the air. During a period of 35 years (1900–1935) it increased from 0.029 to 0.032%. The content of carbon dioxide in the air slowly decreases with increasing height.

Air samples from great heights were taken for the first time in the 30's of the 20th century using stratospheric balloons. Later this was done by means of automatic instruments. It was found that up to heights of 20–25 km the composition of the atmosphere varies only slightly. It appears that the composition of the atmosphere does not vary essentially up to heights of 200–300 km and that even higher it consists mainly of nitrogen and oxygen. The content of oxygen begins to decrease slowly from a height of 18 km, and according to theoretical calculations has a value of about 18% at a height of 29 km as compared to 20.95% at the surface of the Earth.

In addition to the ordinary biatomic oxygen, the stratosphere contains ozone, O_3. Its highest concentration occurs at the layer between 14–16 and 30–35 km, with a maximum at heights of 25–28 km. Above and below this layer the amount of ozone decreases, although it is detectable between 5 and 60 km.

Oxygen undergoes some changes at heights of 90-100 km. Here, under the action of the ultraviolet radiation of the Sun, the oxygen molecules dissociate into atoms, and atomic oxygen, O, appears. Its amount decreases rapidly with height, and at a height of 110 km atomic oxygen constitutes 95.6% of the total amount of oxygen at this layer. There are good reasons to assume that above this level most of the oxygen is atomic. The specific weight of atomic oxygen is half the specific weight of molecular oxygen, O_2.

The spectra of the aurora polaris indicate that in the upper layers of the atmosphere nitrogen is also in atomic state, N.

This information on the composition of the atmosphere at various heights is only recent. In the 19th century only the lowest layers of the atmosphere were accessible to man: Analysis of the chemical composition of air on mountains and in air samples taken with the aid of balloons showed that the composition of the air in the troposphere remains unchanged. The main conclusions on the composition of the upper atmosphere were made on the basis of the well-known gas law, discovered at the beginning of the 19th century by the English scientist Dalton.

According to this law, the total pressure of a mixture of several gases, which do not react with one another chemically, under constant temperature is equal to the sum of the pressures of each gas, i.e., to the pressures which each gas of the given mixture would exert if it occupied the entire volume of the mixture. In other words, each gas is distributed in space independent of the presence of the other gases, forming its own atmosphere.

Since the air pressure and density decrease with increasing height, the pressure of a heavier gas decreases with increasing height faster than that of a lighter one. Hence the conclusion that the composition of the atmosphere varies radically with height, since the specific weights of the gases constituting the air are very different. Calculations showed that being a heavier gas, the amount of oxygen should decrease with height faster than the amount of nitrogen. Therefore, at great heights in the atmosphere nitrogen and other lighter gases – helium and hydrogen – should dominate. Textbooks of meteorology and physics, published in the 20's and even in the 30's of this century, contain diagrams representing the composition of the atmosphere at various heights. According to these diagrams the amount of oxygen decreases rapidly with height and completely disappears at heights of 90-100 km. The amount of nitrogen, on the other hand, increases up to heights of 40–50 km and then decreases, disappearing at heights of 100–120 km. It was also assumed that light gases such as hydrogen and helium dominate at heights of 80–100 km, and that above 120 km they are almost the only ones constituting the atmosphere. Later it was found that these assumptions were wrong, since Dalton's law is inapplicable in the real atmospheric conditions.

It was not known whether a constant disorderly vortex, or turbulent motion, causing mixing of the gases not only at the lower layers, but also at great heights, takes place in the atmosphere. From observations of luminous clouds and the motion of meteoric trails it was found that strong winds, associated with turbulent motions, prevail at high layers of the atmosphere. These prevent the separation of the gases according to their specific weight, and the process of slow, diffusive penetration of some gases into others becomes practically impossible.

Information on the composition of the atmosphere at great heights was obtained in the 40-50's by means of photographs of the spectrum of the aurora polaris. Already in the 30's photographing of the spectra of the aurora polaris, as well as of the twilight and the night sky, showed that in the higher layers of the atmosphere nitrogen and oxygen dominate, and that the amount of light gases, when observed, is negligibly small. Thus, the complete absence of hydrogen was concluded. This problem is, however, still open. As to the other light gas — helium — it is probable that there is very little at great heights. Calculations show that as a result of the disintegration of the radioactive elements in the terrestrial core during millions of years, the amount of helium escaping to the atmosphere is several times the amount observed. It is assumed that helium constantly escapes from the Earth's atmosphere to interplanetary space.

In addition to the above-listed gases, the atmosphere always contains a considerable amount of dust, arriving from the surface of the Earth and from interplanetary space, and also salts, combustion products, and some amount of water vapor.

In contrast to the basic gases, the amount of water vapor decreases rapidly with height, and most of it is contained in the lowest layers of the atmosphere. In addition, water vapor possesses a remarkable property: its amount depends on the air temperature. The higher the temperature, the higher the moisture content of the air, and vice versa. Water vapor reaches saturation when its amount at a given temperature reaches a certain level. Thus, at a temperature of 34°C above zero 1 m^3 of air may contain a maximum of 35 g of water vapor, at a temperature of 1°C above zero, 4 g, and at 25°C below zero, only 0.4 g, Therefore, initially unsaturated air may become saturated when cooled. As soon as the amount of water vapor in the air reaches a certain level, further cooling will transform it into water droplets.

At a low temperature the water droplets usually freeze, transforming into ice crystals. Observations show, however, that even at temperatures of 10°C - 15°C below zero, and sometimes even lower, very small water droplets may remain unfrozen, i.e., stay in a supercooled state. At a very low temperature the water vapor is often transformed at once into ice crystals, avoiding the water phase. This process is called sublimation.

Since in the lower layer of the atmosphere the air temperature falls with increasing height, the amount of water vapor rapidly decreases accordingly. The first 5-kilometer layer contains about 90% of the total amount of water vapor; in middle latitudes almost all the water vapor is contained in the first 10- 12 km. Above 10-12 km the atmosphere is almost clear of water vapor.

The dependence of the water vapor content of air on its temperature plays an important role in the organic life of our planet, since it causes the formation of clouds and precipitation.

The principle dust sources in the lower layers of the atmosphere are the dry land regions. Air masses moving over arid zones of the continents contain a considerable amount of dust. The larger dust particles rapidly drop out and the remaining lighter ones are carried over huge distances suspended by the wind, causing the turbidity of the air and impairment of visibility. Higher layers of the atmosphere contain dust which arrives from interplanetary space (cosmic dust). The air is also contaminated by the action of volcanoes, forest fires, industrial towns. It is enriched by salt admixtures from the seas and oceans. When the water is agitated, water splashes containing salts are carried away by the wind. The water droplets evaporate rapidly, and the very small salt particles remain in the air in a suspended state and are carried to the continent.

All these dust, ash, and salt particles constitute the so-called "condensation nuclei", i.e., nuclei on which the water vapor condenses when the air is saturated, thus contributing to the formation of clouds.

These impurities in the air affect the color of the sky. The molecules of the gases constituting the air, together with the dust particles, water droplets, and ice crystals partially scatter the solar radiation. The gas molecules scatter the short-wave radiation, i.e., the violet rays, blue rays and so on, and therefore by day the sky has a blue color. The other particles contained in the air, which are considerably larger than the gas molecules, scatter radiation of almost all wavelengths. Thus the sky acquires a whitish color in the presence of dust or water droplets. Observations from air balloons show that high up the sky becomes dark-violet, since the amount of dust in the atmosphere decreases with height.

Pressure and weight of the air

It is obvious to us now that air has weight and that its pressure decreases with height, but it was not so obvious 300 years ago. Until the middle of the 17th century air was considered invisible and weightless. It is true that already at the time of Aristotle, in the 4th century B.C., unsuccessful attempts were made to prove that air has weight. However, until the middle of the 17th century people were far from the idea that piston pumps could not operate if air did not have weight and could not exert pressure.

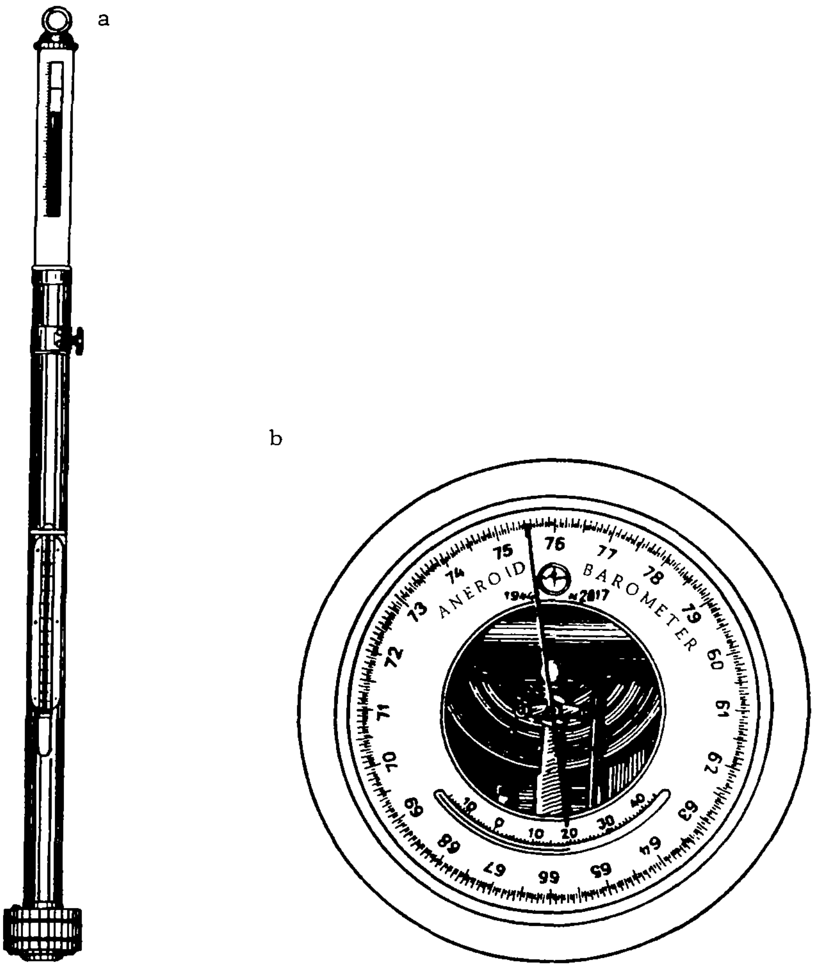

FIGURE 1. Mercury barometer (a) and metallic aneroid barometer (b)

In 1642 Torricelli showed for the first time that a water column 32 ft high (i.e., approximately 10.33 m) balances the atmospheric pressure. Since mercury is 13.6 times as heavy as water, the atmospheric pressure is balanced by a mercury column 76 cm high. A tube, filled with mercury and placed with its open end in a vessel, also filled with mercury, was called a barometer (Figure 1). Four years after Torricelli, similar experiments were performed by Pascal and Ruane (France). Thus, the fact that air can exert pressure was finally established.

If the atmosphere has weight and exerts a pressure, the magnitude of the latter should be different at different heights. Experiments performed by Perier in the September of 1648, using a mercury barometer, showed that the air pressure at the top of a mountain 975 m high is considerably lower than at its foot. It was thus proved that air pressure decreases with increasing height.

Since gases have weight, the atmosphere is held by the force of the Earth's attraction and does not scatter into interplanetary space, but enveloping the Earth, rotates with it. On each square meter at sea level, the air mass exerts a pressure of 10,333 kg. In other words, this is the weight of an air column of cross section 1 m^2 extending from sea level to the upper boundary of the atmosphere. A cubic meter of air at sea level weighs about 1.3 kg.

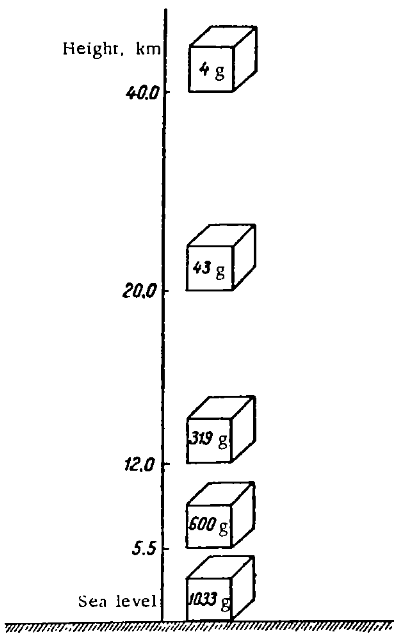

FIGURE 2. The weight of 1 m^3 of air at various heights

If the air density were constant at all heights, the thickness of the air envelope of the Earth would be approximately 8000 m*. However, as is shown by light scattering data, the atmosphere extends to approximately 2000 km. The reason for this discrepancy is that the air density decreases rapidly with height. 1 m^3 of air at a height of 5.5 km already weighs not 1.3 kg but about 600 g, i.e., less than half, and at a height of 40 km — about 4 g (Figure 2). Of two equal volumes of cold and warm air, the latter weighs less.

On the bottom of the air ocean the invisible but perceptible air presses on each square centimeter with a force of 1033 g. It is interesting to determine the pressure to which our body is subjected. If the external surface of the human body averages about 12,000–15,000 cm^2, it is subjected to a load of 12,000–15,000 kg, or 15 tons. However, the organism does not feel this load, since the external pressure is balanced by the air pressure inside the body. Life on the Earth is adapted to just this pressure. Therefore, when ascending to great heights the state of health of man deteriorates not only because of the insufficiency of oxygen, but also because of the considerably rarefied medium under a very low pressure.

* The weight of an atmospheric column of cross section 1 m^2 being equal to 10,333 kg and that of 1 m^3 of air being 1.3 kg, it should be easy to calculate the height of the atmosphere: one would just have to divide 10,333 by 1.3.

Experiments in a pressure chamber have shown that the more complex the organism, the more difficult for it to stand low air pressure. For example, a representative of the cold-blooded animals, the frog, can survive at a height of 20–30 km for several hours, whereas man when rapidly ascending to a height of 7–8 km loses consciousness usually after 8–10 minutes. At heights of 15–16 km, even if man breathes only oxygen, oxygen starvation proceeds after a few seconds due to the low pressure. Under low atmospheric pressure the blood begins to boil. It is known that the lower the pressure, the sooner water boils. On mountains, for example, water boils not at 100°C, but at a lower temperature. At a height of about 20 km water boils at a temperature of 37°C above zero. Blood also boils at this height. Therefore in interplanetary flights man is provided with the appropriate physiological and hygienic conditions. In the cabin of the spaceship, a more or less constant temperature is maintained; the cosmonaut is protected from noise and vibrations, and is provided with a convenient working place, good illumination, etc.

Data obtained from the first and second satellite spaceships launched in the Soviet Union have demonstrated the appropriateness of the design of the cabins provided for man's flight.

The pressure exerted by an air column bounded by the surface of the Earth and the upper boundary of the atmosphere is measured in millimeters of mercury by means of a mercury or aneroid barometer (see Figure 1). The statement "the atmospheric pressure at sea level is equal to 760 mm" means that the air pressure there is equal to the pressure of a mercury column 760 mm (76 cm) high. It is customary to express the pressure in thousandths of a bar, i.e., in millibars*. The mean pressure at sea level, expressed in millibars, is equal to 1013.3. An air pressure of 1000 mb is equivalent to the pressure of a mercury column 750.1 mm high. To convert millimeters into millibars a conversion factor of 4/3 is usually used. Using Table 1, millimeters of mercury can be converted into millibars.

Similar to the decrease in air density with increasing height, the atmospheric pressure also drops rapidly. Therefore, in spite of the large vertical extension of the atmosphere, half its mass is concentrated up to the first 5–6 km. The pressure at this level is only 500 mb, i.e., half that at sea level; 0.9 of the entire mass of the atmosphere is contained in an air column about 16 km high. The pressure at this level is 100 mb, and at a height of 40 km – only 2.4 mb.

The decrease in the atmospheric pressure with height can be characterized by the vertical pressure gradient, or by the so-called barometric step. B a r o m e t r i c s t e p is the distance along the vertical (in meters) at which the atmospheric pressure varies by one unit (by 1 mb). The magnitude of the barometric step varies. It depends on the height above sea level and on the air temperature. In the surface layer under a pressure of 1000 mb and at a temperature of 0°C the barometric step is 8 m. This means that at 8 m intervals the pressure falls by 1 mb. At the 600–500 mb layer, corresponding to heights of 4.5–5.5 km, the barometric step is 13 m, while at the 100-200 mb layer it is 40 m.

* Bar – is the pressure produced by a force of 1 million dynes per 1 cm^2 of area. The dyne is the unit of force in the CGS (centimeter-gram-second) system. A dyne is a force which gives a mass of 1 g an acceleration of 1 cm/sec^2.

TABLE 1

Conversion of millimeters of mercury into millibars

Hundreds and tens of mm of mercury	Millimeters of mercury									
	0	1	2	3	4	5	6	7	8	9
	Millibars									
700	933.2	934.6	935.9	937.3	938.6	939.9	941.2	942.6	943.9	945.2
710	946.6	947.9	949.2	950.6	951.9	953.2	954.6	955.9	957.2	958.6
720	959.9	961.2	962.6	963.9	965.2	966.6	967.2	969.2	970.6	971.9
730	973.2	974.6	975.9	977.2	978.6	979.9	981.2	982.6	933.9	985.2
740	986.6	987.9	989.2	990.6	991.9	993.2	994.6	995.9	997.2	998.6
750	999.9	1001.2	1002.6	1003.9	1005.2	1006.6	1007.9	1009.2	1010.6	1011.9
760	1013.2	1014.6	1015.9	1017.2	1018.6	1019.9	1021.2	1022.6	1023.9	1025.2
770	1026.6	1027.9	1029.2	1030.6	1031.9	1033.2	1034.6	1035.9	1037.2	1038.6
780	1039.9	1041.2	1042.6	1043.9	1045.2	1046.6	1047.9	1049.2	1050.0	1051.9
790	1053.2	1054.6	1055.9	1057.2	1058.6	1059.9	1061.2	1062.6	1063.9	1065.2
800	1066.6	1067.9	1069.2	1070.6	1071.9	1073.2	1074.6	1075.9	1077.2	1678.6

It is convenient to use the barometric step for approximate calculations of the pressure variation with height. Thus, for example, it is easy to determine the difference in the air pressures between the first and twenty-fifth floors of a high building. If the difference between the heights of these floors is 90 m, the air pressure at the 25th floor is approximately 12 mb less than at the level of the first floor. However, we do not feel the pressure difference even when ascending in a high-speed elevator. This is because a pressure variation of about 10–20 mb is very small (1–2%) compared to the normal atmospheric pressure. The mean day-to-day pressure variation to which the human body is accustomed is small. In tropical zones it amounts to about 1 mb, at middle latitudes 5–6 mb. In some individual cases, however, the day-to-day pressure variation in middle latitudes reaches 20–30 mb and more.

Methods of investigating the atmosphere

To study the physical state of the atmosphere, both instrumental and visual observations are carried out. Instrumental observations are conducted at meteorological stations by means of special instruments mounted at ground level and by instruments which are lifted by balloons, airplanes, and kites. Instrumental observations supply information on the temperature, humidity, and pressure of the air, and on the velocity and direction of the wind at ground level and at heights of up to 30-40 km. In addition, the height of the lower and upper boundaries of the clouds, the amount of precipitation, the composition of the air, the distribution of the radiant energy, etc., may be determined.

Visual observations are conducted at meteorological stations (Figure 3). In these observations the form and amount of clouds (i.e., the percentage of covered sky), the range of horizontal visibility (the degree of transparency of the air), the character of atmospheric precipitation, the intensity of snowstorms, etc., are determined.

Indirect methods for studying the structure of the atmosphere also exist. They are mainly used for obtaining information concerning high layers of the atmosphere, which for the time being are hardly accessible for sounding. Indirect methods include observations of light phenomena in the atmosphere, propagation of sound waves and radio waves. Such light phenomena as the aurora polaris, luminescence of the night sky, trail of meteors, brilliance of the twilight sky, and others, make it possible to estimate the density and temperature of the air, and the velocity and direction of air currents.

Among the indirect methods for studying the atmosphere we mention the following:

From mother-of-pearl clouds the wind and air humidity at heights of 22-26 km is determined; from noctilucent clouds - air currents at heights of 80-90 km may be studied.

From the anomalous propagation of sound the temperature, pressure, and wind are determined; these elements are also determined from meteor trails at heights of 50–150 km.

From the ultraviolet radiation the ozone content is determined, while the radiation of the night sky furnishes information on the composition and temperature of the air at heights of 60–70 km; similar information may be obtained by observing the aurora polaris at heights of 80–1000 km.

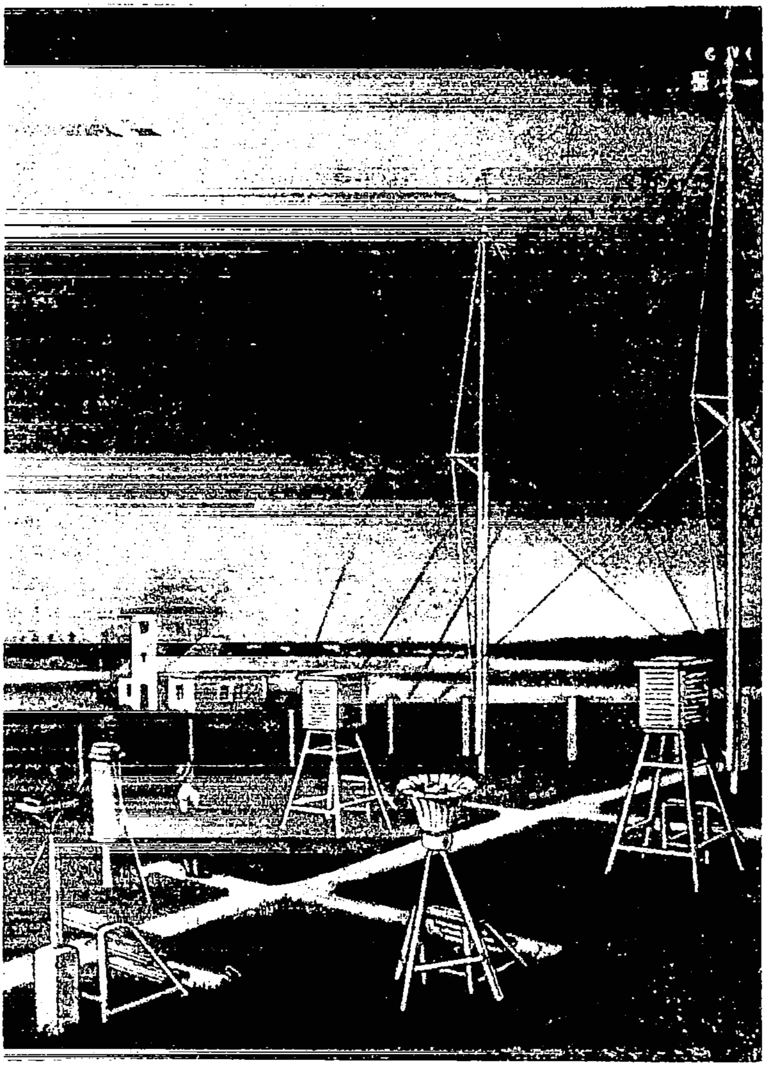

FIGURE 3. Meteorological station

By meteorological and geophysical rockets the pressure, density, and temperature of the air, as well as the solar spectrum, etc., are determined.

The most widespread radiometeorological instrument is the radiosonde – an invention of P.A. Molchanov (Figure 4). Released with a rubber balloon in the free atmosphere, the radiosonde records during its flight the pressure, temperature, and humidity of the air, and transmits the results by radio signals of a certain form. The signals are received by radio receivers and decoded by the observers. After a rapid processing, the values of the meteorological elements at various heights are obtained.

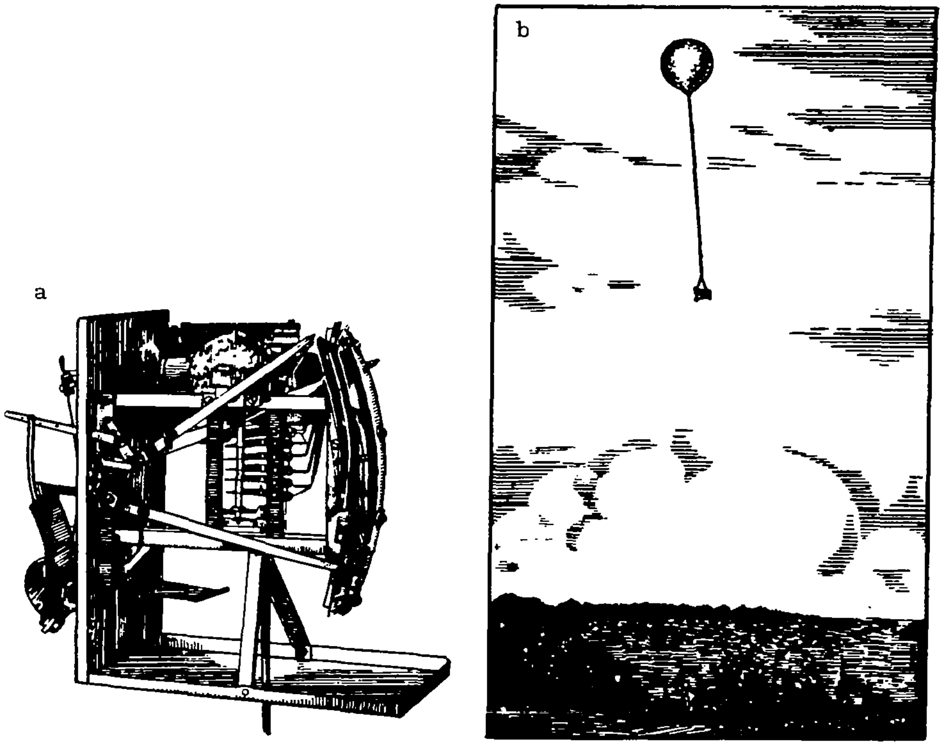

FIGURE 4. Radiosonde (a) and radiosonde in flight (b)

Information on the direction and velocity of air currents high up are obtained by means of pilot balloons and radar pilots. Pilot balloons are small rubber balloons filled with hydrogen. After launching they are observed in free flight with the aid of an aerological theodolite. From angle measurements the direction and velocity of the wind at various heights are calculated. In contrast to pilot balloon observations that are conducted in clear weather, radar pilot observations by means of radar or direction finder make it possible to determine the direction and velocity of the wind also in cloudy weather.

The height of the lower boundary of the clouds is measured by means of pilot balloons and projectors. For this purpose use is also made of airplanes equipped for atmospheric soundings, and of ceiling-height indicators, lifted on rubber balloons. Recently special airplane laboratories were equipped to study the microstructure of clouds as well as for other investigations.

Almost all the above-listed means of observation on the physical state of the free atmosphere were originated in the present century, mainly during the last 20–25 years. Aerology is the branch of meteorology that deals with the study of the physical processes and the phenomena in the free atmosphere.

The first data on the structure of the atmosphere were obtained by means of aerostats. The first scientific application of the aerostat in Russia was made by Academician Ya. D. Zakharov in 1804. Later, flights were made by the well-known scientists D.I. Mendeleev, M.A. Rykachev, and others. The flight of D.I. Mendeleev took place on 7 (19) August, 1877 from the town of Klin.

The first flight of a stratospheric balloon in Russia took place in 1933. The stratospheric balloon "SSSR-1" ascended to a record height for that time of 19 km (Figure 5). Another Soviet stratospheric balloon "Osoaviakhim-1" reached a height of 22 km in 1934. The observations conducted during that flight yielded valuable information on the structure and composition of the air in the lower layers of the stratosphere. Stratospheric balloon flights were also carried out in the USA at this time.

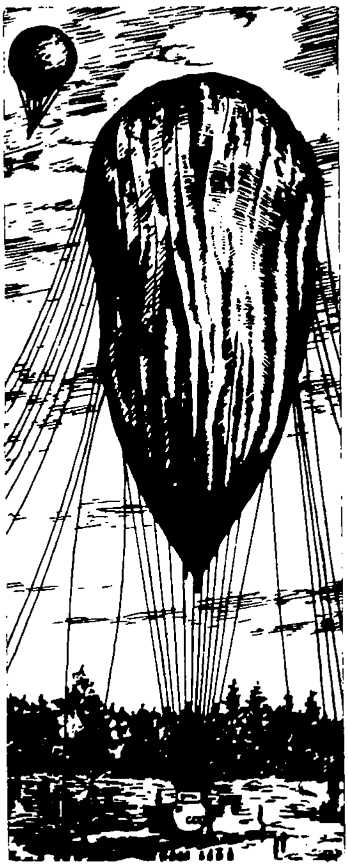

FIGURE 5. The stratospheric balloon "SSSR-1"

New interesting data on the structure of the higher layers of the atmosphere, as mentioned before, were obtained at the end of the 40's and during the 50's by means of special meteorological and geophysical rockets, artificial Earth satellites, and indirect methods of investigation of the atmosphere. Many launches were made during and after the IGY, i.e., from 1957 onwards.

Numerous rocket launches both in the USSR and abroad were made from diverse points of the northern and southern hemispheres. As a result, valuable data on the higher layers of the atmosphere were obtained for the first time over the Arctic and Antarctic, Europe and Asia, America and Australia, and over the oceans. Particularly interesting data were collected in the Arctic, the Antarctic, and the equatorial zone.

Most meteorological rockets are launched to a height of 60–100 km. Geophysical rockets attain considerably greater heights. For example, an equipped Soviet rocket having a total weight of 2200 kg ascended in May, 1957 to a height of 212 km, and on 21 February, 1958 another Soviet rocket with scientific equipment of total weight 1520 kg reached a height of 473 km. The instruments placed in the rocket were usually recovered on Earth.

The recording of various meteorological elements and phenomena takes place during the rapid ascent of the rocket as well as during the smooth descent of the container by parachute with the equipment separated from the rocket engine. The results are transmitted to the ground by means of radiotelemetric equipment. The scientific instruments record the temperature, pressure, and chemical composition of the atmosphere at various heights; the physical properties of the ionosphere, cosmic rays, and the shortwave ultraviolet part of the solar spectrum are also studied in a similar way.

The launching of the first artificial Earth satellite started a new era of investigation of outer space.

Artificial Earth satellites have considerable advantages over meteorological rockets for studying the atmosphere. The latter, being very expensive and complex, only obtain information at a few points, i.e., at their launching and during short time intervals. For a systematic investigation of atmospheric processes a broad network of stations which simultaneously launch rockets, similar to the existing network of aerological stations, is needed: this is very difficult to realize at present.

Artificial satellites, in spite of the difficulty of launching them into orbit, possess a number of advantages. Being a scientific laboratory, a satellite during the course of its many days of flight records and transmits by radio information on the composition of the atmosphere, cosmic radiation, intensity of the magnetic field of the Earth, corpuscular radiation of the Sun, etc., over the entire terrestrial globe at the height of its orbit.

Special meteorological Earth satellites photograph the clouds from a height of 300 km and more, thereby recording the character of the weather simultaneously over vast regions of the Earth. From the data obtained by means of artificial satellites, the components of the thermal balance of the atmosphere are calculated; from this the temperature and wind distribution at ground level and at various heights can be determined.

It is obvious that a number of meteorological artificial satellites can be launched simultaneously to various heights; this would allow repeated and large time-interval data on the features of the processes in the higher layers of the atmosphere to be obtained. In order for an artificial satellite to stay in its orbit for long periods of time its orbit must be situated above the dense layers of the atmosphere, i.e., above 200 km.

Artificial Earth satellites, launched into orbits below 1000 km over the terrestrial surface, pass through the upper layers of the atmosphere. Coming into contact with the atmosphere and subjected to its drag, the satellites gradually lose their velocity and pass to lower orbits. Artificial satellites that are launched into orbits over 1000 km above the terrestrial surface may stay there for a long time.

The first artificial Earth satellite was launched in the Soviet Union on 4 October, 1957 to a height of about 900 km, the second – on 3 November, 1957 to a height of 1700 km, and the third – on 15 May, 1958 to a height of 1880 km.

The launching of cosmic ships increased considerably methods for studying interplanetary space. The first Soviet cosmic ship was introduced into orbit on 15 May, 1960. The launching of the second ship took place on 19 August, 1960, of the third – on 1 December, 1960.

Cosmic rockets are launched to study interplanetary space. The first cosmic rocket with a total weight of 1472 kg was launched in the Soviet Union on 2 January, 1959, the second – on 12 September (its weight was 1511 kg), and the third – on 4 October of the same year (weight 1553 kg).

New ways of penetrating the atmosphere and interplanetary space were found during 1961. On 12 February a rocket was launched in the Soviet Union to the planet Venus, and on 12 April, the first cosmonaut of the world Yuri Alekseevich Gagarin made a flight around the Earth on the "Vostok-1" cosmic ship. The flight, which took 108 minutes, aroused admiration throughout the world.

The date 12 April, 1961 will go down in history as that on which man's penetration into the cosmos began. The historic feat of Yuri Gagarin demonstrated the extent of the creative genius of the Soviet people.

The second cosmic ship, weighing about 4.6 tons, returned safely to Earth. All the conditions for man's flight existed. However, complete reliability and safety of the flight and the return of the cosmonaut to Earth was still lacking. It was only after a series of launchings that Soviet scientists sent the first man on a cosmic flight. Later, the USA launched manned rockets and a satellite into space.

A number of difficulties must be overcome before flights with cosmic ships can be realized. Already in the 17th century the great Newton determined two velocities necessary for overcoming the Earth's force of attraction. One of them – the first cosmic velocity – is equal to 8 km/sec at the surface of the Earth. This velocity transforms a body launched around the Earth into an artificial satellite. The second velocity, called the second cosmic velocity, is equal to 11 km/sec. On acquiring the second cosmic velocity, the launched body overcomes the force of the Earth's attraction and escapes into interplanetary space. These velocities are attained by means of multi-stage rockets.

For a safe manned cosmic flight interplanetary ships should be guided, since in this case return to the Earth can be secured. But this is still not enough. Conditions must be such that the human organism withstands the flight. The human organism can easily support any velocities. We do not feel the velocity of a train, plane, the Earth around the Sun (this velocity is approximately 30 km/sec), etc. But the human organism is very sensitive to velocity variations, i.e., to accelerations. Some people can easily stand riding on "roller coasters", while others feel bad even when ascending and descending in an elevator.

The acceleration of a cosmic ship is tremendous. This increases the weight of the cosmonaut several fold at the moment of takeoff. Therefore, in addition to the special training of the organism for flight in the cosmos, a mode of ascent has to be worked out which takes into account the safety of the cosmonaut.

And what is the effect of weightlessness on man?

In a vertical launching to a height of 100 km man is subjected to weightlessness during approximately 3 minutes, in a launching to 200 km – 5 to 6 minutes, and to 500 km – about 10 minutes. In the orbital flights of artificial Earth satellites, as well as of cosmic ships, weightlessness is maintained continuously.

Experimental flights with animals showed that weightlessness should not affect the organism appreciably. Yuri Gagarin's flight finally settled the effect of weightlessness on man's body.

Less than four months after man's first flight into the cosmos Soviet science achieved a new brilliant success: the realization of cosmic flights.

At 0900 hrs on 6 August, 1961 the Soviet cosmic ship "Vostok-2", piloted by German Stepanovich Titov, completed 17 revolutions around the Earth in 25 hours. Then, having flown over 700,000 km, it landed on 7 August at 1018 hrs in a pre-assigned region, near the landing place of the "Vostok-1" of the first cosmonaut Yuri Gagarin.

The flight of the "Vostok-2" took place in an orbit with a minimum distance from the Earth's surface (perigee) of 183 km and a maximum distance (apogee) of 244 km. The flight proved that man can withstand a prolonged stay in interplanetary space.

On 11 August, 1962 the cosmic ship "Vostok-3", piloted by cosmonaut Andryan Grigorevich Nikolaev, was launched into orbit from the USSR. On the next day another ship, "Vostok-4", piloted by cosmonaut Pavel Romanovich Popovich, was launched into orbit.

The period of revolution of both ships around the Earth was 88.5 minutes. The maximum distance of the ships from the surface of the Earth (apogee) was 251 and 254 km respectively, and the minimum distance (perigee) – 183 and 180 km.

The first group flight in a cosmic ship took place in the ionosphere (thermosphere). Our knowledge about this region is still very limited.

The Soviet ships landed on 15 August at about 0100 hrs. The flight program was executed completely.

The ship "Vostok-3", having completed 64 revolutions around the Earth, traversed a distance of over 2.6 million km in 95 hours, and the ship "Vostok-4", 48 revolutions during its 71 hours flight, having traversed a distance of about 2 million km.

The remarkable flights of the Soviet cosmonauts Yuri Gagarin, German Titov, Andryan Nikolaev, Pavel Popovich, and of the American cosmonaut John Glenn and others showed that man will soon be able to penetrate into interplanetary space and realize the dreams of flight to the Moon and to the planets of the solar system.

The structure of the atmosphere

The atmosphere is inhomogeneous both vertically and horizontally. It is divided along the vertical into a number of layers that differ in their physical characteristics. Along the horizontal, particularly in its lower part, it is made up of inhomogeneous air masses. The air layer nearest to the Earth's surface is called the troposphere.

The troposphere. The physical properties of the troposphere are determined to a large extent by the influence of the terrestrial surface. The lower boundary of the troposphere is the surface of the Earth, the upper boundary is situated on the average at heights of 8–17 km. The height of the troposphere depends mainly on the geographic latitude. Its greatest height is observed in the equatorial zone: here it reaches 16–18 km. Over the near-polar and neighboring regions, the upper boundary of the troposphere lies on the average at a height of 9–10 km. At middle latitudes, the height of the troposphere oscillates from 6–8 to 14–16 km, being on the average 10–12 km.

The upper boundary of the troposphere undergoes seasonal variations: during the winter it is lower, during the summer, higher. The variations in the height of the troposphere are even larger, depending on the character of the atmospheric processes. Often over a period of 24 hours the height of the upper boundary of the troposphere over a given point or region varies by several kilometers. Observations show that the vertical variations of the upper boundary of the troposphere are related to variations in the air temperature.

The troposphere possesses a number of physical properties which differentiate it from higher layers. A considerable part of the mass of the terrestrial atmosphere and almost all the water vapor contained in it are concentrated in the troposphere. In addition, the temperature falls on the average by 0.6°C per 100 m in the troposphere. The air in the troposphere is heated and cooled mainly by the Earth's surface. In accordance with the inflow of solar energy, the temperature drops from the Equator to the Poles. Thus, the mean air temperature at the surface of the Earth at the Equator reaches 26°C above zero, whereas in polar regions it reaches 23°C below zero. At the same time, the temperature over the Equator in the upper troposphere is between -75° and -80°C, whereas in polar regions it is between -60° and -65°C.

The prevailing horizontal air transfer in the troposphere is westerly. The wind velocity in the troposphere generally increases with height, attaining a maximum at the level of its upper boundary. Horizontal transfer is associated with vertical mixing of the air and turbulent motion, providing continuous mixing of the air throughout the troposphere. Due to the ascent and descent of large air masses, clouds are formed and dispersed and precipitation occurs. The processes which determine the weather and its variations develop in the troposphere.

Above the troposphere is the stratosphere. It is separated from the troposphere by a transition region called the *tropopause*.

Like the height of the upper boundary of the troposphere, the height of the tropopause undergoes seasonal and daily variations that depend on the processes which develop in the troposphere. When lying over cold air masses it is situated very low, when over warm masses – high. During the winter, the tropopause is often situated at heights of 8–9 km even at middle latitudes and during the summer — at heights of 13–15 km. The variations in the height of the tropopause are due to a number of reasons. A great factor is the transfer of cold or warm air masses and the cooling or heating of the air caused by its vertical motions. When the temperature in the troposphere rises, the tropopause rises, when the temperature drops, so does the tropopause.

The stratosphere. According to the vertical division of the atmosphere, the stratosphere is that region which is bounded below by the tropopause and above by the 50–60 km level.

The only really great difference between the stratosphere and the troposphere is the temperature distribution with height. The stratosphere contains very little water vapor. No stormy processes of cloud formation, associated with precipitation, take place.

Until quite recently it was believed that the stratosphere is a relatively quiet medium and that no air mixing takes place there in the vertical direction. It was also assumed that the temperature in the stratosphere is determined by the radiation balance alone, i.e., by the equality of absorbed and reflected solar radiation.

New data, obtained by means of radiometeorological instruments and meteorological rockets, showed that in the stratosphere, as in the upper troposphere, intensive air circulation with considerable temperature and wind variations takes place. There, as in the troposphere, considerable vertical mixing, and turbulent motions associated with strong horizontal air currents, are observed. All this is a result of the nonuniform temperature distribution.

Table 2 shows the temperature in the stratosphere as a function of the latitude in the northern hemisphere.

TABLE 2

Distribution of the mean air temperature with height between 10° and 80° NL in January, °C

Latitude, degrees	Height, km						
	5	9	12	16	20	24	30
80	−38, −40	−58, −60	−57, −65	−64, −68	−62, −70	−65, −74	−67, −75
70	−34, −40	−56, −60	−54, −64	−56, −64	−53, −66	−54, −70	−58, −72
60	−30, −40	−52, −58	−52, −59	−52, −61	−46, −62	−44, −61	−47, −64
50	−26, −36	−47, −52	−51, −58	−51, −58	−47, −58	−44, −58	−44, −56
40	−22, −31	−43, −50	−51, −56	−53, −60	−50, −59	−44, −57	−44, −50
30	−12, −17	−37, −43	−52, −57	−64, −69	−60, −63	−50, −60	−45, −49
20	−7, −11	−32, −36	−51, −56	−70, −75	−65, −68	−54, −60	−46, −49
10	−2, −5	−28, −32	−50, −53	−76, −80	−66, −72	−57, −60	−46, −50

We see from this table that at heights of 5 and 9 km in the troposphere the temperature difference between low and high latitudes attains 30 to 35°C, gradually falling from low to high latitudes. In the stratosphere the temperature distribution is somewhat different. At the 16-km level the lowest temperatures (-76° to -80°C) are observed in the equatorial zone, at middle latitudes the temperature is -51° to -61°C, and toward high latitudes falls again to -64° to -68°C. In the equatorial zone of the stratosphere the temperature rises with height attaining -46° to -50°C at the 30-km level, and in the arctic zone temperatures of about -67° to -75°C are observed at this level.

During the summer the temperature distribution undergoes considerable variations. As follows from Table 3, the temperature in the troposphere at 5 and 9 km falls, as during the winter, from low to high latitudes, its

TABLE 3

Distribution of the mean air temperature with height between 10° and 80° NL in July

Latitude, degrees	Height, km						
	5	9	12	16	20	24	30
80	−20, −23	−44, −46	−43, −44	−42, −43	−39, −40	−39, −40	−35, −36
70	−16, −21	−40, −45	−44, −48	−44, −45	−40, −44	−40, −42	−37, −38
60	−14, −18	−37, −43	−48, −50	−47, −50	−44, −48	−44, −48	−39, −42
50	−10, −14	−33, −40	−48, −53	−50, −56	−47, −51	−46, −50	−41, −44
40	−6, −10	−26, −40	−43, −56	−58, −64	−52, −56	−49, −52	−44, −45
30	−1, −8	−23, −36	−43, −57	−64, −72	−56, −60	−51, −53	−45, −46
20	−1, −7	−25, −34	−46, −56	−70, −77	−60, −62	−52, −54	−45, −46
10	−4, −7	−28, −32	−48, −54	−72, −76	−62, −64	−54, −56	−46, −47

difference being about 15°; this is due to the summer heating of the air at middle and particularly at high latitudes. At the 16-km level the temperature rises from the equatorial zone to 80° NL, up to -42° to -43°C, and even at the 30-km level it is higher in the Arctic than in the equatorial zone.

From the given data on the temperature distribution with height in various latitudes, it follows that in the upper layers of the equatorial zone of the stratosphere the air temperature does not vary appreciably from winter to summer, whereas in the Arctic zone these variations are quite large.

Table 4 gives the temperature differences between summer and winter at various levels in the troposphere and stratosphere and at various latitudes in the northern hemisphere.

TABLE 4

Temperature differences in the middle latitudes between July and January at heights of 0–30 km, °C

Height, km	Latitude, degrees									
	0	10	20	30	40	50	60	70	80	90
Ground level	—1	1	6	13	17	24	29	32	31	40
5	1	0	4	11	14	17	18	19	19	17
9	1	0	5	9	13	14	13	18	14	15
12	1	0	3	4	4	4	9	14	17	20
16	4	3	0	—1	—2	2	10	16	22	24
20	7	6	6	3	5	7	14	22	27	30
24	6	4	5	6	5	10	17	25	31	34
30	2	2	3	4	4	10	19	27	38	41

It can be seen from the data of Table 4 that the temperature differences between summer and winter increase from low to high latitudes. At the 30-km level over the polar region they reach a maximum (40°C). The same is true for the southern hemisphere, the only difference being that in the Antarctic the differences in the temperature at this level reach 50° to 55°C.

The mesosphere. Observations by means of meteorological rockets and by indirect methods have established that the general rise in temperature observed in the stratosphere continues up to heights of 50-60 km. At these heights the air temperature rises to 10° to 20°C above zero. Above this layer it falls again and at the upper boundary of the mesosphere (about 80 km) attains a value of -75° to -90°C. Higher up the temperature again rises.

Figure 6 contains curves of the mean temperature as a function of height between ground level and the 90-km level for three latitudes: 80°, 50° and 20°. The curves show the inhomogeneity of the atmospheric structure over these latitudes for any given season. It can easily be seen that *even during the same season and at the same level, temperature differences* between different latitudes exceed 20° to 30° C. The inhomogeneity is particularly marked in the low-temperature layer in the stratosphere (18–30 km), in the maximum-temperature layer in the middle mesosphere (50–60 km), and in the low-temperature layer in the upper mesosphere (75–85 km).

The rather complex system of air currents in the stratosphere and in the mesosphere is caused by the seasonal temperature distribution.

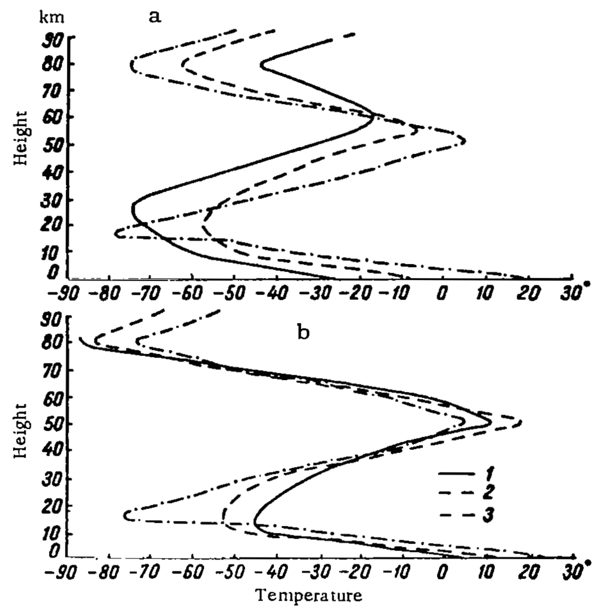

FIGURE 6. Variation curves of the mean air temperature as a function of height between the ground level and the 90-km level for 80° (1), 50° (2) and 20° (3) NL during January (a) and July (b)

Figure 7 contains curves of the mean wind velocity as a function of height between ground level and the 90-km level for the same latitude zones as in Figure 6. The curves show a considerable difference between the wind velocity and direction distribution during January and July. As can be seen, north of 20° NL, westerly winds with mean maximum velocities of over 340 km/hr prevail during January. During July, westerly winds prevail in the troposphere and in the lower stratosphere up to heights of 18–20 km, whereas higher up they change into easterly winds (in the figure these velocities are indicated with a minus sign). In the lower thermosphere the winds again become westerly. During the winter, on the other hand, westerly winds are transformed into easterly winds above the mesopause level.

Clouds are observed at those elevations where the temperature drop with height is replaced by isothermal conditions or by inversion.

At heights of 20-26 km in the upper stratosphere thin and not dense, so-called mother-of-pearl clouds appear under certain conditions (obviously, in the case of sharply pronounced inversions); these clouds consist of ice crystals and supercooled water droplets (Figure 8).

Clouds are also observed at a height of about 80 km, i.e., where the air temperature stops falling and begins rising with height (see Figure 6). If the weather is fine at twilight during the summer brilliant thin clouds brightly illuminated by the Sun from beyond the horizon can be seen under the inversion layer. These are called noctilucent clouds (Figure 9).

It is assumed that noctilucent clouds consist of ice crystals. Like pearly clouds, they apparently arise due to the accumulation of water vapor over the temperature inversion layer (I. A. Khvostikov). The role of condensation nuclei is probably played there by cosmic dust. The level of the noctilucent clouds is obviously determined by the "stopping layer", formed due to temperature increase with height in passing from the mesosphere to the higher lying layer — the thermosphere.

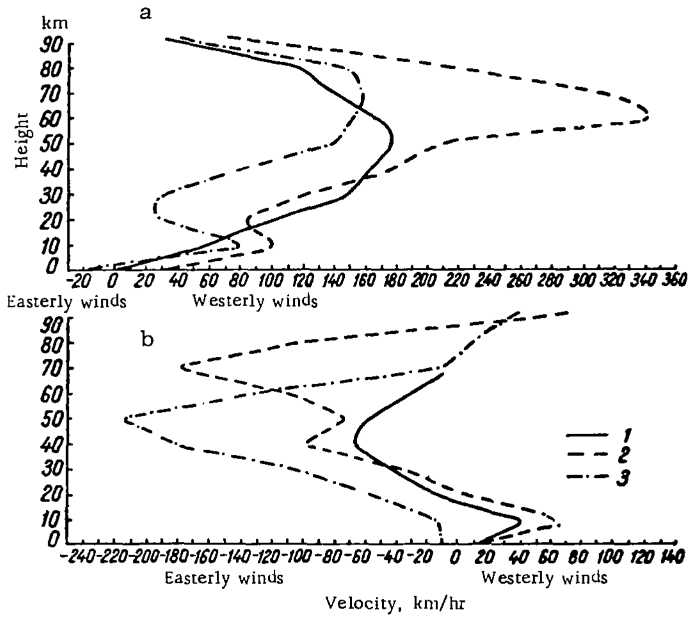

FIGURE 7. Variation curves of the mean wind velocity as a function of height between ground level and the 90-km level for 80° (1), 50° (2) and 20° (3) NL during January (a) and July (b)

Observations of noctilucent clouds revealed that the winds at their level vary considerably during the summer. The wind velocities oscillate within wide limits (from 50 to several hundreds of kilometers per hour).

The thermosphere. Above the mesosphere lies the region called the thermosphere. Characteristic of this region is the temperature rise with height. From data obtained by means of rockets and by indirect methods of temperature determination it has been established that at the 150-km level the temperature reaches 220° to 240°C, and that at the 200-km level it exceeds 500°C. At higher levels the temperature continues to rise and at the upper boundary of the thermosphere, at the 700–800-km level, it exceeds 1000°C. However, for the high layers of the atmosphere the concept of "temperature" has a different meaning.

The temperature of a gas is determined by the mean velocity of its molecules. In the lower, denser part of the atmosphere collisions of molecules are frequent and their kinetic energy is on the average constant. When air molecules absorb a large amount of radiant energy they possess high kinetic energy and this energy is instantly distributed among the molecules. For this reason all molecules possess the same kinetic energy and, consequently, the same temperature.

FIGURE 8. Mother of pearl clouds

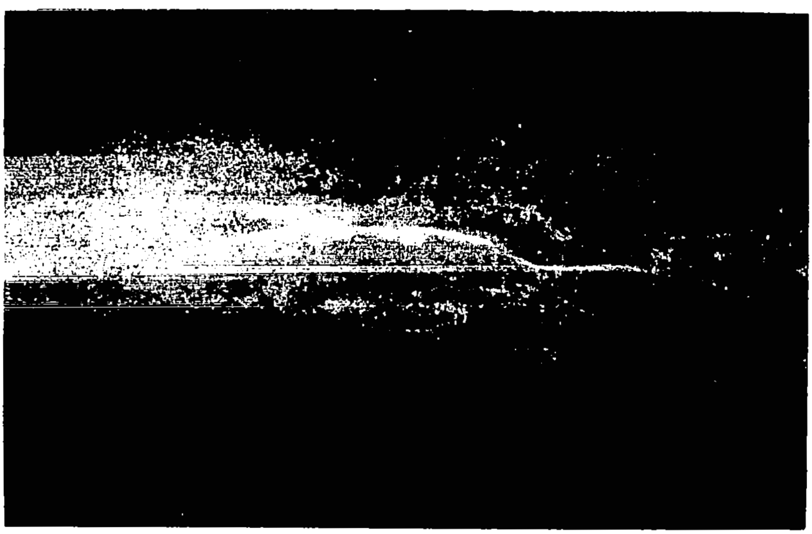

FIGURE 9. Noctilucent clouds

In the high layers of the atmosphere, where the air density is very low, collisions between molecules are infrequent. Upon the absorption of energy the velocity of the molecules in the time interval between their collisions varies greatly. In addition, molecules of light gases have higher velocities than molecules of heavy gases. The temperature of these gases may be different.

The extremely high temperatures in the thermosphere indicate only that in this highly rarefied medium some molecules move with enormous velocities. A body situated there does not even "feel" temperatures of 1000°–2000°C. The main part of the meteorite burns out in the thermosphere, not reaching the surface of the Earth.

The most interesting feature of that part of the atmosphere above 60 km is its ionization, i.e., the presence of a huge number of electrically charged particles – the ions. At those times of maximum concentration of ions, the atmosphere becomes electrically conducting by virtue of the ionization. Since the ionization is characteristic of the thermosphere, it is also called the ionosphere. The air is ionized under the action of the ultraviolet and corpuscular radiation of the Sun.

The ionization process is most intensive in the thick layers between the heights 60 to 80 and 320 to 400 km. In these layers optimum conditions for ionization exist. The air density is appreciably higher than in the upper atmosphere, and the ultraviolet and corpuscular radiation of the Sun is sufficient for the ionization process.

The ionosphere is divided into a number of layers according to the intensity of the ionization process. The E layer is situated at a height of 100 km, the F_1 and F_2 layers – at heights of 150 to 180 and 220 to 400 km, respectively. In the 60- to 80-km layer, i.e., in the upper mesosphere (the D layer), the ionization process is weaker.

A distinctive feature of the ionosphere is its influence on the propagation of radio waves. Radio waves are refracted, reflected, and absorbed in ionized layers.

The D layer extends up to a height of 80 km. There, long radio waves are absorbed more than reflected due to the higher density of this layer. The remaining ionospheric layers (E, F_1, and F_2) reflect mainly medium and short radio waves, particularly the F_2 layer which is situated at the 220- to 400-km level.

The strong absorption of short radio waves in the ionosphere disrupts radiocommunications. This phenomenon is connected with the variation in solar activity. At times solar spots, which are associated with an intensification in the ultraviolet radiation, appear on the Sun. The electron density of the ionosphere and the absorption of radiowaves at daytime hours then increase, leading to the disruption of the normal operation of shortwave radiocommunications. This is because when the solar radiation intensifies, charged particles (corpuscles) are deflected by the magnetic field of the Earth toward high latitudes. Entering the atmosphere, the corpuscles increase the ionization of the gases so much that they begin to glow. This is how the aurora polaris arises, having the form of beautiful multicolored arcs and draperies, displayed in the night sky mainly at high latitudes. The aurora polaris is associated with strong magnetic storms.

By photographing the aurora polaris from two points several tens of kilometers apart, the height of the aurora is determined with great accuracy.

Usually the lower edge of the aurora polaris is situated at a height of about 100 km, the upper part being observed at a height of several hundreds of kilometers, and sometimes at a height of about 1000 km. The reasons for its variety of forms and color combinations etc., are not yet known.

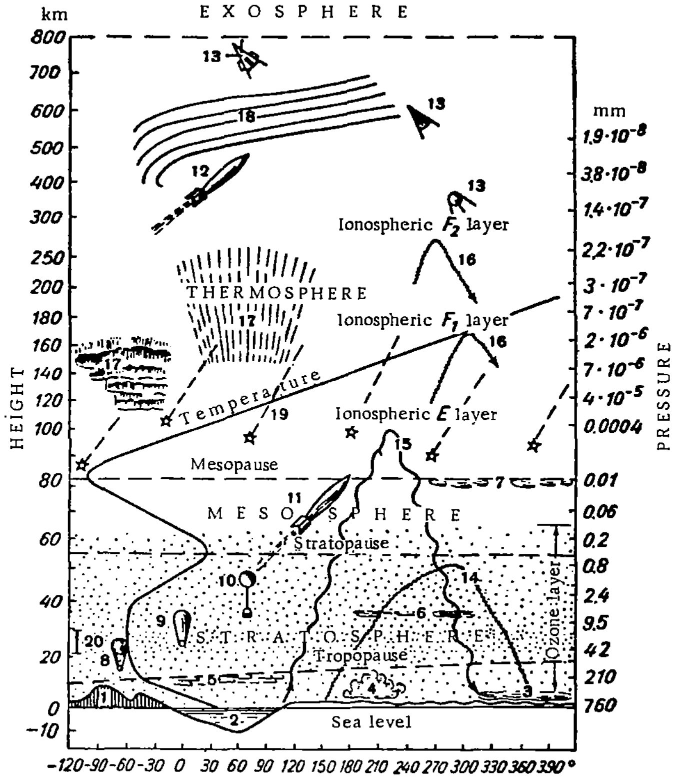

FIGURE 10. Vertical cross section of the atmosphere

1 – greatest mountain height (Everest); 2 – greatest ocean depth; 3 – low-level clouds; 4 – convection clouds; 5 – cirrus clouds; 6 – mother of pearl clouds; 7 – noctilucent clouds; 8 – Picard's stratospheric balloon; 9 – the stratospheric balloon "Osoaviakhim-1"; 10 – radiosonde; 11 – meteorological rockets; 12 – geophysical rockets; 13 – artificial satellites; 14 – reflection of sound waves; 15 – reflection of medium radio waves; 16 – reflection of short radio waves; 17 – aurora polaris in the lower ionosphere; 18 – aurora polaris in the upper ionosphere; 19 – meteors; 20 – layer of highest ozone concentration.

In the case of strong magnetic storms, the aurora polaris becomes visible also at middle latitudes, and in rare cases even in the tropical zone. The intensive aurora, observed on 21–22 January, 1957, could be seen in almost all the southern regions of the USSR.

The ionosphere was sounded for the first time during the 50's by means of rockets and artificial satellites. The processes taking place in the ionosphere were studied also by indirect methods – from the intensity and character of such phenomena as the night glow, the aurora polaris, etc.

The exosphere. This is the highest part of the atmosphere, situated above 800 km. Less is known about it than all the others. Data obtained by means of indirect observational methods and by theoretical calculations show that the temperature in the exosphere rises with height up to 2000°C. In contrast to the lower ionosphere, the gases in the exosphere are so rarefied that their molecules, moving with tremendous velocities, practically never collide with each other.

Until recently it was assumed that the boundary of the atmosphere is situated at a height of about 1000 km. However, from the deceleration of artificial satellites it was found that at heights of 700–800 km, 1 cm^3 contains up to 160,000 positive ions of atomic oxygen and nitrogen. This indicates that rarefied layers of the atmosphere extend to a height of 2000 km and more.

Figure 10 shows a schematic vertical cross section of the atmosphere; the vertical coordinate gives the height and air pressure, the solid curve gives the variation of the air temperature with height. At the corresponding heights, the main phenomena observed in the atmosphere, as well as the maximum heights attained by radiosondes and other means of sounding of the atmosphere, are indicated.

The gaseous tail of the Earth. At the high temperatures that exist at the upper boundary of the atmosphere, the velocities of the molecules reach approximately 12 km/sec. With such velocities the gas particles gradually escape from the Earth's gravitational field into interplanetary space. This process takes a long time. For example, hydrogen particles at a height of about 300 km take several years to escape into interplanetary space, and helium particles — over millions of years. Heavier gases take even more time to escape beyond the limits of the terrestrial atmosphere.

Investigation of the night glow shows that the form of the Earth's air envelope is not spherical: regions of extremely rarefied gases, like the tails of comets, stretch out from the external layers of the terrestrial atmosphere in the plane of the ecliptic to the nonilluminated side of our planet. Judging from the spectrum, the gaseous tail of the Earth consists of oxygen and nitrogen.

The tail is apparently formed as a result of the pressure of the solar radiation on the upper layers of the atmosphere.

Only several years ago it was assumed that beyond the terrestrial atmosphere, in interplanetary space, the gases are highly rarefied and that concentrations do not exceed a few particles per cm^3. We now know that interplanetary space is a relatively dense medium with a concentration of hundreds of particles per cm^3. However, the interplanetary medium, as well as the nature of the gaseous tail of the Earth, is still insufficiently studied.

THE INFLOW OF SOLAR ENERGY

The energy of the Sun

Solar energy is the source of meteorological, hydrological, chemical, biological, and other processes taking place on Earth. All the other incoming energy (radiation of the stars and planets, cosmic rays, internal heat of the Earth, and others) is negligible as compared with that from the Sun. Solar radiation, propagating in space with a velocity of 300,000 km/sec, traverses from the Sun to Earth, a distance of about 150,000,000 km, in 8.3 min (Figure 11). In spite of the huge distance separating us from the Sun and of the position of the Earth in interplanetary space, the surface of the Earth and the lower layers of the atmosphere are sufficiently heated by solar radiation for life on our planet to be possible.

The total amount of energy received by the Earth from the Sun can be compared with the amount of energy generated by the continuous operation of 543 billion steam engines each of 400 hp. This collosal amount of energy which the Earth receives is only a negligible fraction of the radiant energy emitted by the Sun! Calculations show that this fraction is 1 : 2,200,000,000. Most of the solar energy is dispersed into interplanetary space. To get an idea of its magnitude, let us say that it is sufficient to boil all the water in the seas and oceans of the Earth in 1.5 sec!

Where does the Sun get all its energy from? This problem has always attracted scientists. If the Sun was made of a combustible substance its energy would have lasted for only a short time. If the Sun consisted of coal, for example, it would have burned out in 2000–3000 years. It has been found from various sources of data that the Sun emits now approximately the same amount of energy as it emitted billions of years ago. Until only recently people thought that the production of such a huge amount of energy was due to the compression of the Sun. In that case the solar energy would suffice for only several tens of millions of years. The inexhaustibility of the solar energy was explained by modern physics.

The bodies of the universe consist of atoms. The atomic nuclei contain immense energy reserves which are released in nuclear fission. Fission of the atomic nucleus and the release of atomic energy is realized by means of special devices. A different process takes place on the Sun. At a temperature similar to that of the interior of the Sun, reaching according to calculations 40,000,000°C, and under the collosal pressure, the hydrogen and helium atoms, constituting 90% of the mass of the Sun, disintegrate. A continuous transformation process of the atoms of hydrogen into atoms of helium takes place. As a result of this reaction, the nuclear energy is released.

Thus nuclear transformations are the source of solar energy. The formation of 1 g of helium from hydrogen releases 155 billion cal. To do this only a negligible amount of matter — 0.007 g — is expended. For the Sun this loss amounts to 4,200,000 ton/sec. This, however, has a small effect on the "burning" of the Sun, since the hydrogen reserves on the Sun are so huge that there will still be sufficient for many billions of years.

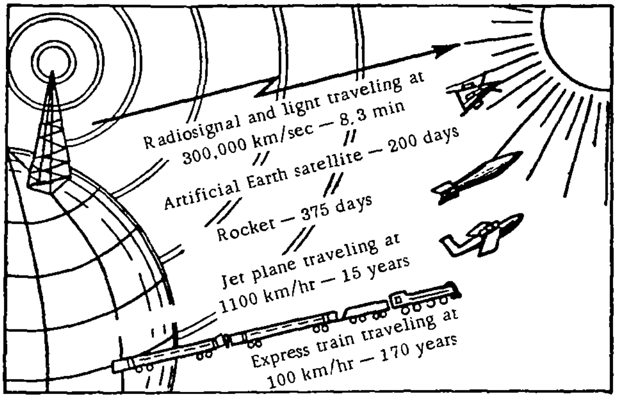

FIGURE 11. Time needed to travel from the Earth to the Sun

The arrival and reflection of solar energy

To study the processes caused by the inflow and loss of solar energy, one must know how much is received by the Earth, the atmosphere, and the hydrosphere, how this energy is distributed over the terrestrial sphere, and how it is spent. The Earth not only receives solar heat but also radiates it back. *The difference between the inflow and loss of the solar radiation energy is called* the radiation balance.

The heat loss from the terrestrial surface by radiation is largely compensated by the radiation from the atmosphere back to Earth. However, because the temperature of the terrestrial surface is higher than that of the atmosphere, the radiation from the Earth is always larger than the radiation of its atmosphere. In other words, the terrestrial surface always loses some heat. The difference between the amount of terrestrial radiation and the amount of the atmospheric counter-radiation absorbed by the underlying surface is called the effective radiation. The terrestrial radiation is measured in calories per square centimeter per minute ($cal/cm^2 . min$).

The temperature of the terrestrial surface and the humidity of the air exert a considerable influence on the intensity of the effective radiation, which is therefore higher during the day than at night, and similarly higher during the summer than the winter. In those regions of the Earth where cloudiness is often observed, the effective radiation of the terrestrial surface is lower than in places where clear weather prevails.

The heat lost by the terrestrial surface is thus considerably reduced owing to the absorption capacity of the atmosphere. With no atmosphere, a different equilibrium would be established between the inflow of solar heat and the radiation from the surface of the Earth.

Already 80 years ago, A.I. Voeikov wrote: "... One of the most important jobs for the physical scientists of today is to keep a record of the inflow and loss of solar heat received by the terrestrial sphere and its air and water envelope. We have to know how much solar heat is received at the upper boundaries of the atmosphere. How much is spent on heating the atmosphere and on changing the state of the water vapor mixed in it. What amount reaches the land and water surfaces. What amount is spent on heating various bodies and on changing their state (from solid to liquid and from liquid to gas). How much is spent on chemical reactions, particularly associated with organic life. We also have to know how much heat the Earth loses by radiation into space, and how this loss takes place ..."*.

The science of the inflow and loss of solar radiation is called actinometry.

Measuring the intensity of solar radiation

The unit of measurement of the intensity of solar radiation is the amount of heat in calories** which 1 cm^2 of surface perpendicular to the incoming radiation receives in 1 min ($cal/cm^2.min$). As shown by theoretical calculations using numerous observations, the intensity of the solar radiation above the atmosphere is about 2 $cal/cm^2.min$. This quantity is called the solar constant. The value of the solar constant undergoes variations that depend on the distance between the Earth and the Sun. The motion of the Earth around the Sun is not along a circle, but along an ellipse. The Sun is located at one of the focuses. During one year the distance between the Earth and the Sun varies. The Sun is nearest to the Earth on 3 January, and farthest away on 3 July. When the Earth is nearest to the Sun the solar constant is 1.94 $cal/cm^2.min$, and when it is farthest away — 1.82 $cal/cm^2.min$. The mean value of the solar constant is 1.88 $cal/cm^2.min$.

The intensity of the solar radiation, as characterized by the solar constant, undergoes some very small variations, depending on the variation in the number of Sun spots.

The amount of heat received by the Earth from the Sun

If there was no atmosphere and if each square centimeter of the terrestrial surface would be perpendicular to the solar rays and would receive 1.88 cal/min at the mean distance of the Earth from the Sun (almost 150 million kilometers), then during one year it would receive about 1000 kcal

* Voeikov, A.I. Klimaty zemnogo shara, v osobennosti Rossii (Climates of the Terrestrial Globe, in Particular of Russia). —Izdatel'stvo Akademii Nauk SSSR. 1948.

** A calorie is the amount of heat needed to raise the temperature of 1 g of water by 1°C. 1000 calories = = 1 kilocalorie (large calorie).

of energy. Since the Earth is spherical and the solar rays do not fall perpendicularly everywhere and since only half of the terrestrial sphere is illuminated, then during one year the upper boundary of the atmosphere receives on the average only a quarter of the above-quoted value, i.e., about 250 kcal/cm^2. The surface of the Earth and the atmosphere absorb up to 140–150 kcal/cm^2 of this solar energy per year.

The amount of heat received from the Sun by the terrestrial surface depends first of all on the angle of incidence of the solar rays*. The more perpendicularly do the solar rays fall, i.e., the higher the Sun over the horizon, the shorter the path of the solar rays in the atmosphere (Figure 12) and the larger the amount of energy arriving per unit area. Conversely, the smaller the angle of incidence, the longer the path traversed by the solar rays in the atmosphere and the less the energy falling on unit area.

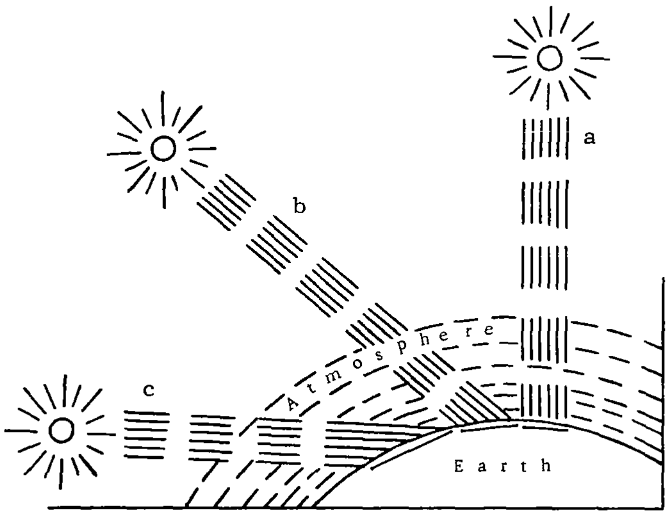

FIGURE 12. The longer the path, the more energy lost by solar rays in passing through the atmosphere. The same beam falling perpendicularly (a) illuminates a smaller area than when it falls at a small angle (b and c)

The maximum amount of solar radiation reaches the surface of the Earth, perpendicular to the solar rays, when the Sun is at the zenith, i.e., when the incident angle of the solar rays is equal to 90°.

In Table 5 calculated amounts of solar radiation for the summer and winter solstice in the absence of the atmosphere are given. It follows from the data of this table that in the absence of the atmosphere the Arctic would receive heat at the rate of 1110 cal/cm^2·day during days of summer solstice, i.e., more than the equatorial zone, where the daily amount of heat would be only 814 cal/cm^2.

Calculations show that in the case of the so-called ideal atmosphere (absolutely dry and pure) the surface of the Earth at high and even at middle latitudes would receive more heat during the summer than the equatorial

* The angle of incidence of the solar rays is the angle formed by the solar ray and the tangent to the surface of the Earth: it is usually identified with the height of the Sun over the horizon.

zone. According to calculations, toward the end of June, in the absence of clouds and with a mean atmospheric transparency, the amount of heat arriving at the North Pole would be about 670 cal/cm^2.day, at 55° latitude 630 cal/cm^2.day, and in the equatorial zone only about 500 cal/cm^2.day.

TABLE 5

Daily amounts of solar radiation in the absence of the atmosphere (northern hemisphere), cal/cm^2.day

	Latitude, degrees									
	0	10	20	30	40	50	60	70	80	90
Summer solstice . .	814	900	964	1005	1022	1020	1009	1043	1093	1110
Winter solstice . .	869	756	624	480	327	181	51	0	0	0

In the equatorial zone the amount of solar heat does not undergo large seasonal variations (Table 5). At the same time, at middle latitudes it decreases considerably, and at the North Pole the inflow of heat completely stops during the period September-March.

During the polar summer the Sun does not set below the horizon, and during winter does not appear above it for a number of days. This causes the latitudinal variation in the solar radiation. In the equatorial zone, however, the duration of the lighted part of the day does not vary appreciably during the year and is approximately 12 hours. Thus low latitudes receive more heat annually than middle and high latitudes.

To find out to what extent the amount of energy arriving on a surface perpendicular to the solar rays depends on their angle of incidence, we turn to Table 6, compiled by N.N. Kalitin. This table gives calculated data on the amount of solar radiation arriving at a perpendicular surface as a function of the height of the Sun over the horizon in the complete absence of an atmosphere (the solar constant), when the solar rays pass through an ideal atmosphere, as well as data obtained from observations in the real atmosphere with mean transparency.

TABLE 6

Theoretical and observed solar radiation intensities for various heights of the Sun above the horizon, cal/cm^2.min

Height of Sun above the horizon, degrees	Solar radiation intensity		
	with no atmosphere (the solar constant)	with an ideal atmosphere	with a real atmosphere
5	1.88	1.05	0.39
10	1.88	1.27	0.65
15	1.88	1.39	0.82
20	1.88	1.47	0.95
30	1.88	1.57	1.11
40	1.88	1.62	1.21
50	1.88	1.65	1.27
60	1.88	1.66	1.31

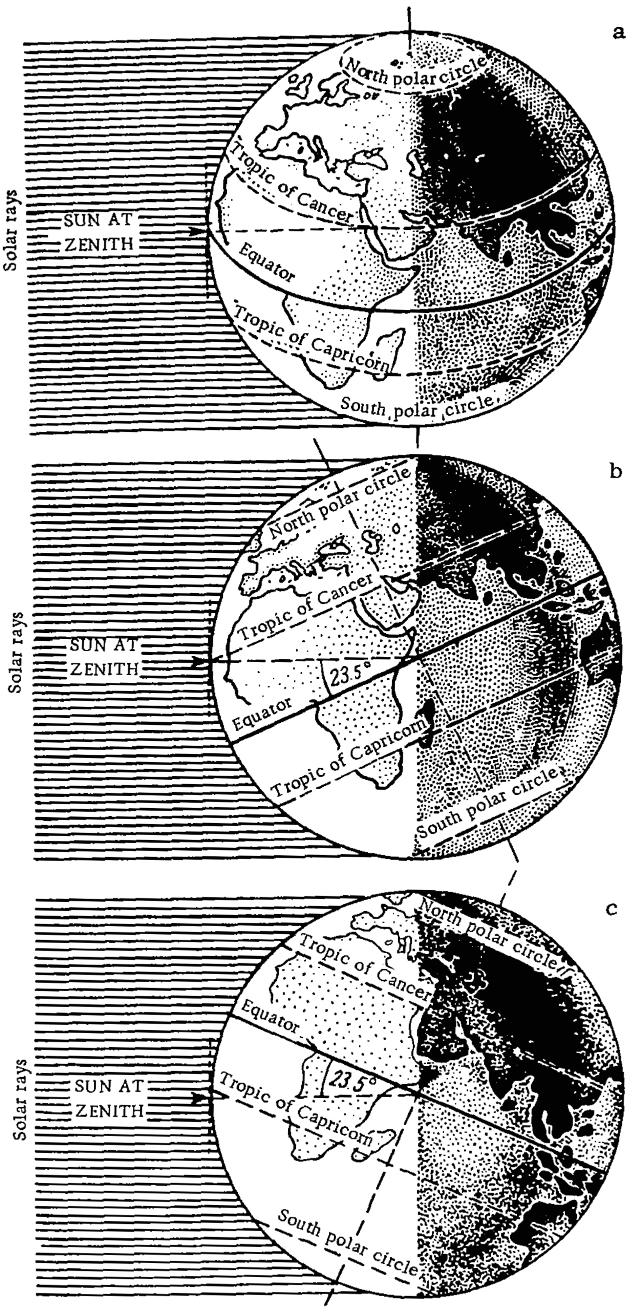

FIGURE 13. Position of the Earth with respect to solar rays for various seasons

a – during the day of the spring and fall equinox; b – during the day of the summer solstice; c – during the day of the winter solstice.

As can be seen from Table 6, even in conditions of an ideal atmosphere the radiation intensity is considerably lower when compared with the solar constant. It is even lower still in conditions of a real atmosphere. With the Sun at 20°, the intensity of the solar radiation is almost half the solar constant, and for a height of 60°, only 30%. The sharp decrease in the solar radiation intensity in the real atmosphere is mainly due to the water vapor and dust that it contains, which absorb quite strongly.

Much less solar energy arrives at a unit of horizontal surface. Thus, when the solar rays fall at an angle of 30°, the amount of radiation arriving on 1 cm^2 of horizontal surface is half that of the data of Table 6, and when the height of the Sun is 5° – almost a twelfth. The solar radiation falling on a horizontal surface rapidly decreases from the Equator to the Poles.

During the spring and fall equinoxes, the Sun is at the zenith at midday on the Equator, whereas at the Poles it is on the horizon (Figure 13 a).

In the summer solstice in the northern hemisphere, the height of the Sun at the Equator is 66.5°, at the Tropic of Cancer 90°, and at the North Pole only 23.5°. At this time the Sun does not set below the horizon in the Arctic and the Polar Day begins. The Antarctic is imbedded in the Polar Night (Figure 13 b).

During the winter solstice in the Arctic the Sun is below the horizon (Polar Night), whereas the Antarctic has the Polar Day (Figure 13 c). However, both at the North and at the South Poles during the Polar Day, the solar rays fall at the smallest angle. The duration of the Polar Day, as well as of the Polar Night, is approximately half a year. Therefore, at low latitudes, where the height of the Sun during the whole year is the maximum, it is much warmer than at middle and particularly at high latitudes of the northern and southern hemispheres. This explains the maximum heating of the terrestrial surface at midday, when the solar rays fall on it at the maximum angle.

Direct and scattered solar radiation

If the atmosphere would let through all the solar radiation, then the climate of any point would depend only on the geographic latitude. This was once assumed to be the case. However, when solar rays pass through the terrestrial atmosphere, as we have already seen, they are weakened by absorption and scattering processes. Water droplets and ice crystals in the clouds are very efficient at absorbing and scattering solar radiation.

That part of the solar radiation which arrives at the surface of the Earth after being scattered by the atmosphere and by the clouds is called the *scattered radiation*. That part of the solar radiation which passes through the atmosphere unscattered is called *direct radiation*.

Radiation is scattered not only by clouds, but also by the gas molecules and by the dust particles. The ratio between the direct and scattered radiations varies within wide limits. Whereas in a clear sky, when the solar rays fall perpendicularly, the fraction of the scattered radiation is 0.1% of the total radiation, in a cloudy sky the amount of scattered radiation may exceed that of direct radiation.

In those parts of the world where clear weather prevails, for example, in Central Asia*, the main source of heating of the terrestrial surface is the direct solar radiation. In places where cloudy weather prevails, as for example, in the north and northwest of the European territory of the USSR, scattered solar radiation becomes important. Let us see what amount of direct and scattered solar radiation various parts of the terrestrial surface receive according to the data of N.N. Kalitin (Table 7). Tikhaya Bay situated in the North, receives almost one and a half times as much scattered radiation as direct. In Tashkent the scattered radiation amounts to less than 1/3 of the direct radiation. The direct solar radiation in Yakutsk is greater than in Leningrad. This is because in Leningrad there are more overcast days and the air transparency is lower.

TABLE 7

Amount of heat recived from direct, scattered, and total solar radiation by 1 cm^2 of horizontal surface, $kcal/cm^2 \cdot year$

Point	Latitude	Radiation		
		direct	scattered	total
Tikhaya Bay	80°19' NL	21	35	56
Yakutsk	62 01	54	27	81
Leningrad	59 41	40	36	76
Irkutsk	52 16	60	30	90
Voronezh	51 40	58	41	99
Tashkent	41 20	103	33	136

Table 8 presents data on the total solar radiation. It can be seen that the annual amounts of total radiation considerably increase from high to low latitudes. From Uedinenie Island to Fresno (California), which are 40.8° latitude apart, the total radiation increases by a factor of almost 2.7.

TABLE 8

Annual amounts of total radiation according to N. N. Kalitin, $kcal/cm^2 \cdot year$

Point	Latitude, degrees	Total radiation	Point	Latitude, degrees	Total radiation
Uedinenie Island	77 NL	64	Stockholm	59.4 NL	76
New Siberian Islands .	73	82	Sverdlovsk	56.9	89
Tiksi Bay.	72	70	Minsk	53.9	83
Saratov	52	108	Warsaw	52.2	91
Karlsruhe	49	98	Nice	44	148
Paris.	49	98	Toronto	44	92
Evpatoriya	46	122	Chicago	42	90
Venice	45	108	New York	41	95
Karadag	45	118	Washington	39	122
			Fresno (California) . . .	37	170

* [Central Asia — A vast region of elevated desert plain and highlands in the interior of Asia, China and the Mongolian People's Republic (MPR). Territory about 6,000,000 km^2. The northern and western borders approximately coincide with the state border between the USSR on the one hand and China and MPR on the other. The southern border runs along the longitudinal tectonic basin of the upper Indus-Tsahgpo River (Beahmaputra). The eastern — along the southern part of the Khingan Mountains, the mountains in the great (Ordes) bend of the Hwang Ho River and the Szechwanese Alps.]

It can be seen from Table 8 that there is no rigorous dependence of the amount of total radiation on the latitude. Thus, for example, Saratov, which is situated 2°–3° to the north of Karlsruhe and Paris, has a larger amount of total radiation than these points. Similar differences can be observed between New York and Nice, Chicago and Evpatoriya, etc.

Albedo of the Earth's surface

The terrestrial surface reflects rays falling on it. The amount of absorbed and reflected radiation depends on the properties of the surface. The ratio of the amount of radiant energy reflected from the surface of a certain body to the amount of radiant energy falling on it is called the albedo of the body. The albedo characterizes the reflective capacity of the surface of the body. When, for example, it is said that the albedo of freshly-fallen snow is 80–85%, it means that 80–85% of the total radiation falling on the snow surface is reflected back.

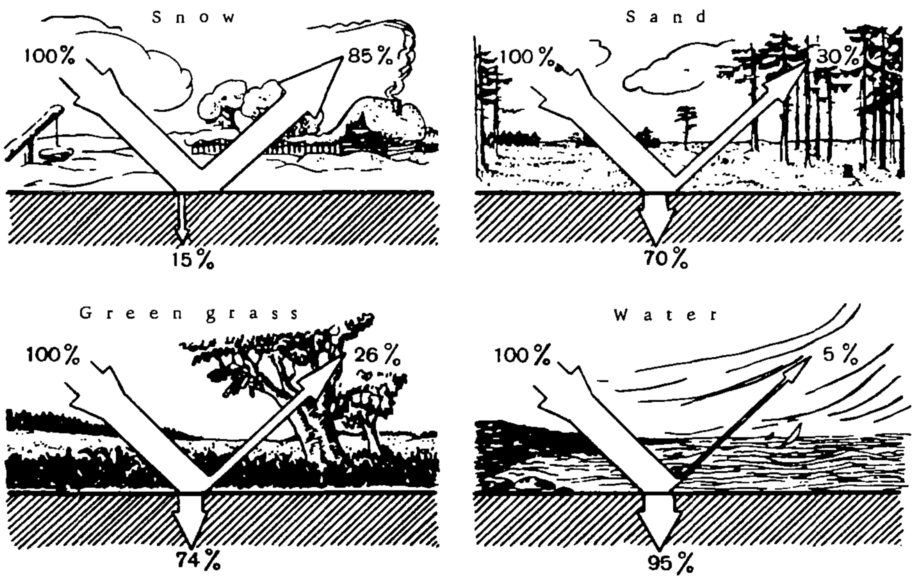

FIGURE 14. Reflecting and absorbing capacity of various surfaces

The albedo of snow and ice depends on their purity. In industrial towns due to the deposition of various impurities, mainly soot, on the snow the albedo is lower. In Arctic regions the albedo of the snow often reaches 94%. Since the albedo of snow is higher than that of other types of surfaces, the heating of the terrestrial surface is weak in the case of a snow cover. The albedo of grass and sand are considerably lower. The albedo of grass is 26%, and that of sand – 30%. This means that grass absorbs 74% of all the solar energy (and sand absorbs 70%); this is used for evaporation, growth of plants, and heating.

Water possesses the highest absorbing capacity (Figure 14). Seas and oceans absorb about 95% of the solar rays falling on their surface, i.e., the albedo of water is 5%. V.V. Shuleikin established that the albedo of water depends on the angle of incidence of the solar rays. When the rays fall perpendicular to the surface of pure water, only 2% is reflected, and when the Sun is low — almost all the rays are reflected.

Long-wave and short-wave radiation and its role in the heat exchange between the Earth and the atmosphere

That part of the radiant energy which reaches the surface of the Earth is spent on heating this surface and the lower layers of the atmosphere. The radiation of the Sun and the Earth are very different. The direct, scattered, and reflected radiation of the Sun, having a wavelength range of from 0.17 to 4 microns*, is called short-wave radiation. The heated surface of the Earth radiates in the wavelength range of from 2 to 40 microns depending on its temperature. This type of radiation is called long-wave radiation. In general, both the radiation of the Sun and the radiation of the Earth contain all wavelengths; but the main part of the energy (99.9%) is contained in the above-indicated wavelength ranges. The difference in the wavelength of the radiations of the Sun and of the Earth plays an important role in the thermal nature of the surface of the Earth.

Heated by the solar rays, our planet itself becomes a source of radiation. Since the temperature of the terrestrial surface does not exceed several tens of degrees, it emits long-wave infrared rays.

Depending on the wavelength, the thermal rays emitted by the terrestrial surface either pass unhindered through the atmosphere or are stopped by it. It has been established that the radiation of 9 to 12 microns escapes freely into interplanetary space; the surface of the Earth loses a considerable part of its heat in this way.

The inflow-outflow of radiant energy, absorbed and radiated by the atmosphere, i.e., the radiation balance, is measured by special instruments, called balance meters.

During the last ten years our knowledge of the thermal balance of the surface of the Earth and of the atmosphere has broadened considerably. There now exist data on the inflow and loss of solar radiation in all parts of the globe. At the Central Geophysical Observatory im. A.I. Voeikov, under the direction of M.I. Budyko, maps of the components of the thermal balance for various months and for a year have for the first time been plotted for the whole globe. They show the thermoenergetic interaction of the surface of the oceans and of the land with the atmosphere. At middle and high latitudes the inflow of heat in summer is positive, in winter — negative. According to calculations of Simpson, as well as of Budyko, the balance of radiant energy is positive during the whole year to the south of 39° NL.

At about 50°N on the European territory of the USSR, the balance is positive from March to November and negative during the three winter

* 1 micron = 0.001 mm.

months. At 80°NL a positive radiation balance is observed only in the May-August period.

In addition to the radiation balance of the underlying surface, defined as the difference between the inflow and outflow of radiant energy of the Sun, the thermal balance of the underlying surface is also calculated.

After absorbing radiant energy from the Sun, the terrestrial surface loses a part of it by radiation. The remaining part is spent on heating the soil and the atmosphere, and on evaporation. Calculating the components of the heat inflow and outflow is an important part of the calculation of the thermal balance of the underlying surface. The basic components of the thermal balance are: the radiation balance, heat exchange in the soil, heat exchange with the atmosphere (or turbulent heat exchange), and the heat lost on moisture evaporation.

Each square centimeter of the globe receives on the average about 0.5 cal/min. Of the total amount of arriving solar energy (100%), 42% is reflected back into interplanetary space, of which 38% is reflected by the atmosphere, and 4% by the surface of the Earth. The remaining 58% is distributed as follows: 14% is absorbed by the atmosphere, and 44% – by the active surface of the soil. Of this, about 18% is lost on evaporation, about 6% on heating of the air, and 20% is lost by the effective radiation of the terrestrial surface.

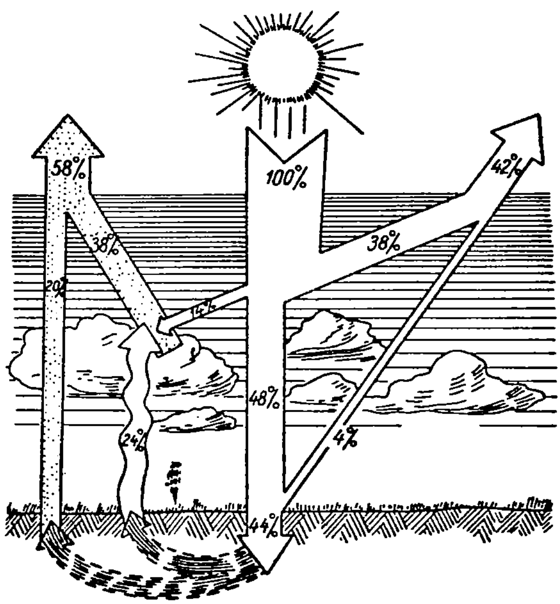

FIGURE 15. Diagram of the thermal balance of the Earth

Thus, of the total amount of solar radiation arriving, the atmosphere receives 14% directly, 6% from the heating of the terrestrial surface, and 18% by the release of latent condensation heat. Together with the effective radiation, amounting to 20%, all the arriving solar radiation (58%) is radiated back into interplanetary space.

Figure 15 gives a diagram of the thermal balance of the Earth, indicating the inflow and outflow of solar energy*.

Data on the inflow and outflow of solar heat are used by meteorologists to explain the complex circulation of the atmosphere and of the hydrosphere, heat and moisture rotation, and many other processes and phenomena taking place in the air and water envelope of the Earth.

The role of water vapor and ozone in the absorption of the terrestrial and solar radiation

The greenhouse effect

Water vapor, ozone, and carbon dioxide absorb the terrestrial radiation. The main mass of water vapor is contained in the troposphere. Of the remaining atmospheric gases at middle latitudes, water vapor constitutes only about 1%; however, by virtue of its large absorbing capacity it absorbs about 25% of the terrestrial radiation. Similar amounts of terrestrial radiation are absorbed by ozone, contained in small amounts in the atmosphere. The strongest absorption by water vapor is that of radiation in the 5.0–7.5 micron wavelength range, by carbon dioxide – in the wavelength range of 12.9–17.1 microns, and by ozone – in the wavelength range of 9.4–9.8 microns. Because they absorb about half the long-wave terrestrial radiation, water vapor, ozone, and carbon dioxide protect the Earth from rapid cooling, creating the so-called "greenhouse effect", which will be discussed below.

Ozone absorbs not only long-wave (thermal) radiation of the Earth, but also ultraviolet radiation of the Sun. The lower layers of the atmosphere contain only a negligible amount of ozone. The amount of ozone in the atmosphere increases with height, reaching a maximum at heights of 24–28 km in the stratosphere, and disappearing at heights of 60–70 km. The ultraviolet rays of the Sun are absorbed to a large extent by the upper layers of the atmosphere. The remaining part penetrates into the mesosphere and stratosphere, and is stopped there by the ozone. The absorption of ultraviolet radiation of the Sun is very important for the development of living organisms, since large doses of ultraviolet radiation kill them. The small doses of ultraviolet radiation passing through the ozone to the Earth's surface, although having a fatal effect on microorganisms, are beneficial to man.

Due to the greenhouse effect, the cooling of the terrestrial surface at night is considerably reduced. It is known that during a cloudless night the air temperature due to the radiation of the terrestrial surface drops considerably compared with the daytime temperature; in these cases a large daily amplitude is observed (by "amplitude" we mean the difference between the maximum and minimum temperatures). During the spring and fall, when the sky clears up at night, the temperature of the air and of the Earth's surface often drops to negative values, i.e., frosts are formed.

A dense cloud cover at night reduces the drop in temperature. For the same evening temperature, for which on a clear night frosts are observed,

* Shifrina, E.M. Solnechnyi luch i ego prevrashcheniya (The Solar Ray and Its Transformations). — Leningrad, Gidrometeoizdat. 1953.

on an overcast night the minimum temperature does not drop to negative values. In this case the greenhouse effect is felt. Part of the heat absorbed by the clouds radiates into interplanetary space, and part is returned to Earth (Figure 16). In the absence of clouds, heat escapes freely into interplanetary space as a result of which the terrestrial surface and the adjacent air layer rapidly cool down.

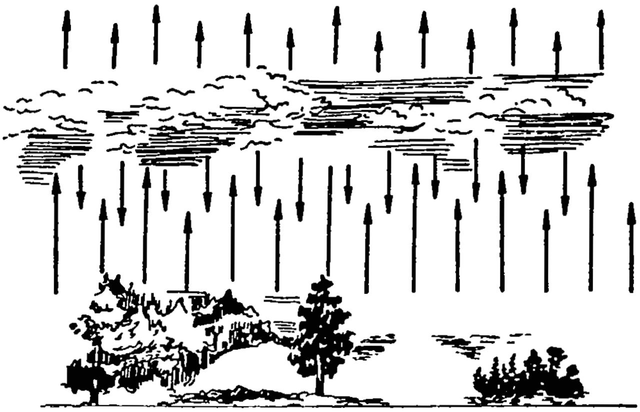

FIGURE 16. The greenhouse effect

The greenhouse effect makes the air temperature at the surface of the Earth equal on the average to 16°C above zero. If we take into account that in interplanetary space the temperature is close to absolute zero, i. e., to -273°C, then the atmosphere makes the temperature at the surface of the Earth 289° higher than the interplanetary temperature.

The greenhouse-effect principle is used for protecting plants from frosts. To fight frosts, smudging is usually applied. The smoke, spreading above the ground, stops the heat emitted by the terrestrial surface.

The principle of the greenhouse effect is used when constructing glasshouses for the cultivation of various heat-loving cultures, for the drying of fruits, etc. Glass has the property of letting through short-wave solar rays and stopping long-wave heat rays. As a result of the accumulation of heat, the temperature inside the greenhouses is considerably higher than the air temperature outside.

The heating of the Earth's surface by solar energy is a different process. It depends to a large extent on the state and thermal conductivity of the underlying surface. Soil, heating rapidly during the day cools down rapidly at night due to the loss of heat by radiation. Dry soil, heating rapidly, transfers most of the heat to the air. Humid soil, and soil covered by vegetation, loses part of the heat by evaporation and plant growth. Solar rays penetrating into a body of water heat not only the upper, but also deeper layers, since in water basins constant mixing of the surface and underlying layers takes place. Because of the mixing, a layer of several tens of meters of the water in seas and oceans heats up during the warm part of the year. Seas and oceans therefore act like accumulators of solar energy. Heat accumulation is particularly high at middle and high latitudes. There the water heats up during the period of the positive radiation balance (during the warm half-year); during the winter, i.e., the period of negative balance, the heat is gradually lost, mainly on heating the air. Consequently in coastal regions during winter the climate is moderate and such strong frosts, as exist in places quite far from seas and oceans, do not occur. For this very reason the mean temperature during January at the southern coast of Iceland is 0°C and at the same latitude in Yakutia it is 48°C below zero.

The use of solar energy

The author is of course aware of the fact that at the moment solar energy is far from being effectively used for domestic and technical purposes, although, as we saw, the amount of solar energy reaching Earth is collosal. According to the calculations of B.P. Vainberg, the founder of rural helioengineering, the amount of solar energy that is technically usable in the USSR may amount to several billions of kilowatts (Table 9).

TABLE 9

Amount of useful solar energy

Region	Technically usable solar energy, billions of kilowatts	Number of months per year
European territory of the USSR	2.7	4
Kazakh SSR	3.2	7
Kirghiz SSR, Tajik SSR, Turkmen SSR, Uzbek SSR	1.4	12
Remaining parts of the Asian territory of the USSR	0.5	3

To get an idea of the figures given in the second column of Table 9, it suffices to say that 1 billion kilowatts is equal to the power of 2000 Dnieper hydroelectric stations. The Central Asian Republics, as can be seen from the table, can use the solar energy during the whole year as can the Transcaucasian Republics.

Large-scale industrial use of solar energy has many inherent difficulties. The main difficulty is that to use a large amount of radiant energy quite large installations are needed. One has also to take into account that solar energy arrives only during the day and depends on the cloudiness, which may affect the operation of the helioinstallations. However, there now exist small helioinstallations, which are mainly for domestic use.

Helioinstallations work on the principle of the greenhouse effect. A glass container made of a thermoinsulating material serves as the receiver of the solar energy. The solar rays penetrate through the receiving glass surface into the container and heat its bottom and walls, which begin to radiate long-wave heat rays. Since glass does not let through long-wave radiation, the heat gradually accumulates in the container. If the receiving surface of the container is covered not by one, but by several glasses with air layers between them, the heating will be more intensive. This is the principle of heaters which raise the temperature to 225°C. Such installations can be used for the operation of baths, laundries, dryers of tobacco, cotton, and fruits, for greenhouses, and for many other purposes.

Scientists from many countries are at work on the problem of the use of solar energy for domestic and industrial purposes.

In India solar stoves, equal in power to an electric stove of 350 w, are being used. In Italy solar pumps are already in operation. Work is being carried out in England, Japan, and other countries. In the Soviet Union, in the USA, and in a number of other countries, conferences are often held for discussing new projects for the use of solar energy.

Several years ago a design of a solar electric station was made at the Power Institute im. Krzhizhanovskii of the Academy of Sciences of the USSR. It is proposed to construct this station in the Ararat valley (Armenia), in the region of the Aigirlich lake. This place was chosen because the number of sunlight hours reaches 2600 per year, i.e., more than in Tashkent and even in Ashkhabad. The output of this unusual electric station is 1200 kwatt. At the center of a huge circular area a 40 m high tower with a rotating boiler will be erected. Around the tower 23 rails will be laid, along which 1300 reflecting mirrors will move automatically, according to the motion of the Sun; these mirrors will direct the solar rays to the boiler with water. The vapor formed under a pressure of 30 atm will drive the powerful turbine of the solar electric station.

In order that the station will work in overcast weather, particularly during the winter months, the design provides for the creation of simple heat accumulators in the form of underground water reservoirs – gigantic heat reservoirs where the water being heated by the Sun will have a temperature of 70°–80°.

This project of a solar electric station aroused great interest all over the world. However, we assume that the application of such cumbersome installations for the use of solar energy will be limited.

The method of indirect transformation of solar into electric energy, avoiding the steam boilers and generators, is more promising.

It is already possible to create energy on a small-scale by means of semiconductor converters. Solar energy will be transformed into electric energy in two ways: thermoelectric and photoelectric.

By the thermoelectric method, the solar energy is directly transformed into electric energy by means of semiconductor thermoelements.

By the photoelectric method, the solar energy is transformed into electric energy by means of semiconductor photoelements (solar batteries). A flux of light falling on the surface of a photoelement creates an electric current.

Semiconductor solar batteries for supplying power to operate scientific equipment and radiotransmitters were used for the first time on the third Soviet artificial satellite.

The first experiments using semiconductors are encouraging. It will apparently not be too long before solar batteries will find a wide application not only for domestic purposes but also in industry.

One of the pressing problems of modern science is the problem of photosynthesis. It is known that plants on absorbing solar energy, assimilate carbon dioxide and create organic material, while releasing oxygen. Plants are the source of life on Earth. According to approximate calculations, plants absorb daily 150 billion tons of carbon dioxide, combine them with 25 billion tons of hydrogen and release 400 tons of oxygen. The green leaf has the capacity to transform solar into chemical energy. In other words, the plants on Earth, by absorbing the radiant energy of the Sun, produce a huge amount of organic material. At the same time, plants assimilate only a small fraction (about 1%) of the absorbed solar radiation. If it was possible to discover the secrets of photosynthesis and increase the assimilation of absorbed radiation by several percent, the productivity of agriculture would rise considerably.

Lecturing at the Academy of Sciences of the USSR, the outstanding physicist and fighter for peace Frédéric Joliot-Curie said: "Although I believe in the future of atomic energy and am convinced of the importance of this invention, I nevertheless think that the real revolution in energy will only come about when we will be able to carry out mass synthesis of molecules, similar to chlorophyl but of even higher quality. To achieve this we must first study in detail this type of molecule and the nature of photosynthesis".

This problem is investigated by scientists all over the world and particularly by the Soviet scientists A.N. Ternin, N.M. Sisakyan, A.A. Nichiporovich, et al.

THE AIR TEMPERATURE AT THE SURFACE OF THE EARTH

Mean air temperature

The amount of heat arriving from the Sun depends on a number of factors: the incidence angle of the solar rays which in turn depends on the geographic latitude, on the time of year, and on the day; the degree of pollution of the atmosphere and its water vapor content; and the absorption and reflection properties of the soil. Nevertheless, in the equatorial zone and in general at low latitudes the amount of solar energy arriving is considerably more than at middle and particularly at high latitudes.

In the equatorial zone each square centimeter of surface receives during a year approximately 32,000 cal more than it radiates back. In the extreme south of the USSR, at approximately 38°NL, the inflow of radiant energy is equal to the radiative loss. At middle and particularly at high latitudes of the northern and southern hemispheres the loss of radiant energy exceeds on the average its inflow. Thus, between 60° and 90° NL, each cm^2 of the terrestrial surface loses more than 60,000 cal per year.

If the air temperature at each point of the terrestrial surface was determined only by the inflow of radiant energy, then its distribution with latitude on the surface of the northern hemisphere would be as shown in the first column of Table 10.

TABLE 10

Calculated mean radiation-balance temperature and actually observed temperature for various latitudes, °C

	Latitude, degrees									
	0	10	20	30	40	50	60	70	80	90
Calculated temperature	39	36	32	22	8	−6	−20	−32	−41	−44
Actual temperature	26	27	25	20	14	6	−1	−9	−18	−20
Difference	−13	−9	−7	−2	+6	+12	+19	+23	+23	+24

In reality, the distribution of the mean temperature with latitude, as can be seen from the second line of this table, is completely different. The differences (third line) between the calculated and actual (radiation-balance) mean temperature vary between −13°C at the Equator and +23°C at 70° or 80°. In other words, at the Equator the mean temperature is 13°

lower than it should be from the inflow of radiant energy, and at latitudes of 70° or 80° the mean temperature is 23° higher. The reason for these differences is the constant interlatitudinal air exchange, which brings hot air from the equatorial zone and from the tropics to middle and high latitudes and, conversely, sends cold air from high and middle latitudes to the tropics and to the equatorial zone.

If no continuous air exchange between low and high latitudes existed, then, as follows from Table 10, there would be very high temperatures at the Equator and in the tropics and on the other hand, at high and middle latitudes very low temperatures would exist. In particular, at the latitude of Leningrad the annual mean temperature would be 19° lower than the observed one.

Thus, the main factors in temperature formation and, moreover, in weather and climate formation are solar radiation, the underlying surface, and the atmospheric circulation.

The weather and climate on Earth are very variable. The meteorological elements that characterize both the climate and the weather are: the air temperature and humidity, the cloudiness and precipitation, the wind, and various phenomena, as for example, storms, snowstorms, fogs, etc. However, weather and climate are not identical concepts. Whereas weather in a given locality is characterized by the state of the atmosphere at any moment, the climate is the weather regime averaged over many years.

The weather at any point on Earth varies continuously. Particularly sharp are the variations at middle and high latitudes. The climate variations of a given locality are less noticeable. Before acquainting ourselves with the climate variations in early times, as well as with its present oscillations (see the chapter "Climate Fluctuations"), we will examine the climatic peculiarities of various countries.

Of the many meteorological elements which characterize the climate we use mainly the temperature and precipitation which give a sufficiently good idea of the peculiarities of a climate. We shall discuss in more detail the climate of the northern hemisphere, learn about the peculiarities of the temperature distribution at the surface of the Earth there, and then describe the variation of the temperature with height.

Peculiarities in air temperature distribution over the terrestrial globe

Climatic variations exist not only along the meridians from the Equator to the Poles, but, as one gets farther from the ocean coasts into the interior of the continents, also along the latitudes. Due to the difference in the amounts of inflowing solar energy, the annual mean-temperature at the Equator and in the tropics is considerably higher than at middle and high latitudes. This difference would have been larger if there were no atmospheric circulation and ocean currents, transferring heat from low to high latitudes.

In addition to the inflow of solar energy and to the heat exchange, the air temperature during any season also depends on the underlying surface, since the degree of heating and of cooling of the air depends on the character of this surface.

Before discussing certain peculiarities of the air temperature distribution in various parts of the terrestrial globe, we should acquaint ourselves with data on the annual mean air temperature at various latitudes. This is shown in Table 11, which was compiled by V. Gorchinskii with corrections introduced from observations in the Antarctic.

TABLE 11

The mean-temperature for January, July, and for a year at various latitudes, °C

Latitude, degrees	January	July	A year	Annual amplitude, degrees
North Pole	- 40.0	- 1.0	- 19.0	39.0
80 NL	- 32.2	2.0	- 17.2	34.1
70	- 26.9	7.2	- 10.4	34.1
60	- 16.4	14.0	- 0.6	30.4
50	- 7.7	18.1	5.4	25.8
40	4.6	23.9	14.0	19.3
30	13.8	26.9	20.4	13.1
20	21.8	27.3	25.0	5.5
10	25.4	26.1	26.0	1.7
Equator	25.3	25.3	25.4	0.6
10 SL	25.2	23.6	24.7	2.2
20	25.3	20.1	22.8	5.2
30	22.6	15.0	18.3	7.6
40	15.3	8.8	12.0	6.5
50	8.4	3.0	5.3	5.4
60	2.1	- 9.1	- 3.4	11.2
70	- 3.5	- 23.0	- 13.6	19.5
80	- 21.0	- 39.5	- 30.2	- 18.5
South Pole	- 25.0	- 48.0	- 36.5	- 23.0

Looking at this table one observes a series of interesting features. For example, it is known that the amount of solar radiation arriving at the terrestrial surface is distributed approximately equally between the northern and southern hemispheres. However, the temperatures at equal latitudes of these hemispheres are considerably different.

Table 12 contains calculated temperature differences between the northern and southern hemispheres, averaged along latitudes.

TABLE 12

Average temperature differences between the northern and southern hemispheres at various latitudes, °C

Season	Latitude, degrees								
	10	20	30	40	50	60	70	80	90
Winter in both hemispheres	1.8	1.7	- 1.2	- 4.2	- 10.7	- 7.3	- 3.9	7.3	12.0
Summer in both hemispheres	0.9	2.0	4.3	8.6	9.7	11.9	10.7	12.8	13.8

Analyzing the data of Table 12, it can easily be seen that between the 30° and 70° latitudes it is appreciably colder during the winter in the northern than in the southern hemisphere. During the summer, on the other hand, it is much warmer over the entire northern hemisphere than over the southern hemisphere. The largest temperature differences, both during winter and summer, are observed at middle and high latitudes. This result, quite unexpected at first, can be explained quite easily if we examine a map: in contrast to the southern hemisphere, land predominates in the northern hemisphere.

Table 13 gives data on the area occupied by land in both hemispheres. Of special importance is the fact that at middle latitudes of the northern hemisphere land constitutes 45–61% of the terrestrial surface, and in the southern hemisphere – only about 4%. The differences at high latitudes are also considerable.

TABLE 13

Percentage of land in the northern and southern hemispheres at various latitudes

Latitude, degrees	Northern hemisphere	Southern hemisphere
90	0	100
80	20	100
70	53	71
60	61	0
50	58	2
40	45	4
30	43.5	20
20	31.5	24
10	24	20
0	22	22

We already know that the continents have a capacity to heat up and to cool down rapidly. Due to the motion of water oceans accumulate tremendous heat reserves during the summer half year but lose them during the winter.

Since the continents in the northern hemisphere occupy a considerable part of the Earth's surface, their rapid heating during the summer and cooling during the winter affect the values of the latitude-averaged temperatures. Thus it is natural that at middle latitudes in the southern hemisphere, where water surfaces predominate, the latitude-averaged winter temperatures are higher than those of similar regions in the northern hemisphere. During the summer, on the other hand, due to the rapid heating of the continents of the northern hemisphere, the air temperature near ground level is considerably higher there than in the southern hemisphere. Thus, the temperature differences between low and high latitudes reach 11–14% there. During the winter too the air temperature is 7–12% higher at high latitudes in the northern hemisphere than at similar latitudes in the southern hemisphere.

Daily air temperature variation

In the absence of horizontal transfer (advection) of heat, as well as of air transformation (see below), the daily variation of the air temperature depends on the following factors: inflow of solar radiation, effective radiation, moisture evaporation, turbulent heat exchange between the underlying surface and the atmosphere, and heat exchange in the soil. Under the various geographic conditions all these factors act differently, changing the daily temperature variation in various parts of the terrestrial globe. Temperature varies with latitudes, over water surface and over land, over humid and dry soil, with cloudiness, with air transparency, etc.

As the Sun rises the inflow of solar radiation increases and heats up the land surface. The heating continues until equilibrium is established between the heat inflow and the heat loss. Because of turbulent heat transfer, the lower and consequently also the higher air layers are heated by the underlying surface. The soil temperature drops at night as a result of reradiation, but this is somewhat reduced by the turbulent heat transfer from the air to the soil. Since the air is heated and cooled by the underlying surface, the greatest difference between the maximum and minimum temperatures is observed near ground level.

With increasing height the daily temperature amplitude decreases. According to the data of S.A. Sapozhnikova, the temperature during the day over the Arys' station (Central Asia) at a height of 5 cm reached 38.1°C above zero in clear weather and at a height of 1.5 m — only 35.2°C for a mean wind velocity of 2.9 m/sec. At night, on the other hand, at a height of 5 cm the temperature dropped to 15.3°C, whereas at a height of 1.5 m with a mean wind velocity of 1.6 m/sec it was 16.3°C. At the 5-cm level the daily amplitude reached 22.8°C, and at the 1.5-m level — only 18.9°C.

In the surface air layer, where the wind velocities are low, the temperature amplitude is larger than at higher layers having higher wind velocities due to weaker turbulent mixing.

The large number of factors influencing the daily time and space temperature variation greatly complicate the prediction of the daily temperature. However, a number of methods exist which make it possible to predict the daily temperature, although only for a few days ahead. M.E. Shvets succeeded in allowing for the influence of the cloudiness and heat losses due to evaporation and drew graphs for determining the daily temperature; these are in operational use in the weather services in the USSR.

Methods for determining the minimum night temperature, based on somewhat different principles, were proposed by D.L. Laikhtman, M.E. Beryland and A.S. Zverev. These methods yield sufficiently accurate values and are used in forecasting fogs and night frosts.

The mean daily amplitude of the air temperature over land in the equatorial and tropical zones reaches 12°-15° C. As one moves northward (at middle latitudes of the northern hemisphere) the daily amplitude decreases. Thus in the southern regions of the USSR during the summer the mean daily temperature amplitude if the sky is clear is 16°-19°C, in the intermediate region 13°-15°C, and in the northern region, 8°-10°C. During the winter the daily amplitude under the same conditions is appreciably lower. In the southern regions of the USSR, for example, it is 10°-12°C.

Over a humid soil or one covered with vegetation, the daily temperature amplitude is smaller than over a bare and dry soil. The daily temperature amplitude depends to an even greater extent on the cloudiness. In an overcast condition, particularly in windy weather, the daily amplitude is small. Moderate or strong winds mix the air, thus reducing its heating and cooling. With an overcast sky the daily temperature amplitude during the summer in southern regions of the USSR is 10°–12°C, in the intermediate region 8°–9°C, and in the north, 4°–6°C.

TABLE 14

Monthly differences between the mean values of the air temperature for a clear sky and overcast conditions at various places in the USSR, °C

Point	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Bratsk	2.1	7.6	10.6	8.4	7.3	8.0	7.1	8.9	9.2	9.6	3.9	-1.0
Igarka	1.0	1.7	3.5	5.9	5.0	3.8	3.3	6.7	7.2	2.9	0.6	0.2
Kazan'	1.8	3.4	5.0	4.2	5.1	5.2	6.0	6.0	7.7	6.7	3.4	1.3
Krasnovodsk	1.4	2.5	2.9	2.7	3.2	2.3	1.6	1.5	3.8	1.5	4.2	3.8
Kushka	8.1	8.3	8.7	9.1	8.5	8.6	9.8	—	13.0	10.1	11.2	10.3
L'vov	5.3	5.0	5.7	6.5	6.2	4.9	8.2	8.8	7.1	7.5	4.9	5.3
Tashkent	6.1	5.2	5.8	5.4	5.9	5.9	5.6	6.4	9.4	8.1	6.7	6.4

The differences in the daily temperature for a clear sky and overcast conditions, according to the data of D.A. Ped' and Z.L. Turketti, are given in Table 14. It follows from the table that in the southern region of the USSR (Kushka) the differences are larger during the year (8°–13°C) than in other regions, and vary slightly from month to month. In the north of the country (Igarka) these differences are somewhat smaller (2°–7°C). At all other stations the differences between the amplitudes during the year vary between the differences at the above-indicated southern and northern stations. At moderate latitudes during the summer these differences are somewhat larger than during the winter. Particularly interesting is the Caspian Sea (Krasnovodsk), in which the differences do not exceed 2°–4°C for almost the entire year.

When the sky is clear the maximum daily temperatures reach high values. Table 15 gives the mean and maximum daily amplitudes for a clear sky at a number of places.

As can be seen from Table 15, in regions with a pronounced continental climate, the maximum values of the daily temperature amplitudes reach 29°–30°C (Bairam-Ali, Oimyakon), whereas at coastal stations they do not usually exceed 10°–12°C (Vladivostok, Kanin Nos, Makhachkala).

In the Central Arctic and Antarctic there is almost no daily amplitude during the year, with the exception of the equinoctial period, due to the fact that during the summer the Sun does not set at all below the horizon, and during the winter it does not rise. The daily air temperature amplitude over a water surface does not usually exceed 2°–3°C, since the temperature of the water surface of seas and oceans undergoes various small variations over a period of 24 hours.

The small daily air temperature amplitude over water basins is due to the fact that the temperature variations of the water surface are transferred by turbulent mixing to lower-lying water layers. According to some data, the daily temperature variations in water reservoirs penetrate to depths of 15–20 m. The penetration depth of the daily temperature variations in soil is much less, about 0.5–1.0 m.

In soil and water the annual temperature oscillations are different. In seas and oceans they penetrate to 60, 200, and even 300 m, whereas in land, only to 15–20 m. For example, in the region of Kaliningrad the annual temperature amplitudes in the sea and in the land are as follows: at the surface of the sea and land the amplitude is 19.0°C and 20.3°C respectively; at a depth of 5 m - 18.6°C and 3.9°C; at a depth of 15 m - 7.5°C and 0.1°C; finally, at a depth of 23 m the annual temperature amplitudes in the sea are 6.5°C whereas in the land the temperature does not vary at all.

TABLE 15

Mean and maximum daily air temperatures for a clear sky in August, °C

Point	Daily amplitude	
	mean	maximum
Bairam-Ali	17.1	29.2
Bramek	17.6	23.9
Dzharkent	19.1	25.5
Kalmykovo	17.4	25.5
Katpa-Kurgan	20.1	29.4
Oimyakon	24.4	30.0
Vladivostok	9.7	11.4
Kanin Nos	6.4	10.9
Makhachkala	7.5	11.8
Fort-Shevchenko	9.2	17.0

Curves of the daily temperature variation at a number of points in clear weather in August are given in Figure 17. As can be seen, a larger amplitude is observed in Yerevan and in Dyushambe, where the day heating

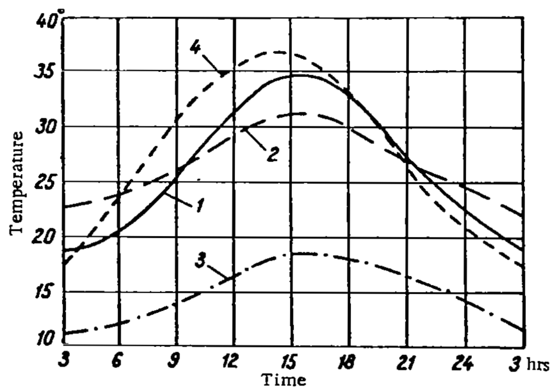

FIGURE 17. The daily air temperature variation in Yerevan (1), Sochi (2), Tallin (3), and Dyushambe (4), in clear weather during the summer

and the night cooling are most intensive because these points are situated in a crater surrounded by mountains. The smallest amplitude is in Sochi, situated on the coast of the Black Sea, and in Tallin, situated on the coast of the Gulf of Finland.

Air temperature in the equatorial zone

In the equatorial zone the mean annual temperature is 25.4°C, the mean temperatures for January and July are both 25.3°C, and the temperature difference between the cold and the warm months is only 0.6°C (see Table 13).

With increasing latitude, the annual variation of the temperature increases. At 40° NL the difference between the mean temperatures for January and July already reaches 19.3°C, and at the latitude of Leningrad (60° N) it reaches 30.4°C. At the Poles this difference is still larger (35°-36°C).

In the equatorial zone the annual temperature amplitude is small not only when averaged over the latitudes. For example, at Manaos (South America), situated approximately 350 km to the south of the Equator, the difference between the temperatures of the warmest (September) and coldest (January) months amounts to 2.2°C, as can be seen from Table 16.

TABLE 16

Monthly mean air temperature at Manaos, Darwin (southern hemisphere), and Kolon (northern hemisphere), °C

Point	Latitude	Longitude	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Manaos	3°07' S	60°02' W	26.0	26.7	26.5	26.6	26.7	26.7	27.0	27.6	28.2	28.2	27.9	27.0
Kolon	9 22 N	79 55 W	26.4	26.2	26.5	26.6	26.6	26.6	26.7	26.3	26.4	26.1	26.1	26.4
Darwin	12 30 S	130 50 E	28.0	28.0	28.5	28.0	26.5	25.0	23.0	25.0	27.5	29.0	29.0	28.5

At Colon (Panama) situated 1000 km to the north of the Equator, the annual temperature amplitude does not exceed 0.6°C. At Darwin (Australia), situated 1400 km to the south of the Equator, the mean temperature of the warm months (October and November) is 20.0°C, and of the coldest month (July), 23.0°C; this gives a temperature difference of 6.0°C.

In Honolulu (Hawaiian Islands), situated in the tropical zone of the Pacific Ocean, the mean January temperature is 21°C, and the mean July temperature, 23°C. During the period of systematic observations, the air temperature there did not fall below 11°C and did not rise above 32°C.

In the Philippine Islands, with an annual mean temperature reaching almost 27°C, the temperature difference between January and May is 2°-4°C. However, in the mountains, at a height of 1.5 km, the mean temperature of the warmest month – May – is only 19°C. Such temperatures are also observed in countries situated between the Equator and the tropics of the American continent. On the territory of Salvador, for example, the annual mean temperature reaches 23°-24°C, and the temperature difference between the cold and warm months does not exceed 3°C.

The southwestern part of the Arabian Peninsula, which is situated at about the same latitude as the Philippines and Salvador, has a much higher annual mean temperature. For example, at Hodeida, situated on the coast of the Red Sea, the annual mean air temperature is 32.5°C, the mean temperature for February reaches 28°C, and for August, 37°C.

These high air temperatures are a result of the strong heating of the land and of the small amount of precipitation. In the coastal zone of the Red Sea the annual precipitation does not exceed 100 mm. However, not far from the coast, on the mountain plateau 2000 m high, the temperature is considerably lower and the annual precipitation exceeds 1000 mm. For example, in Sanaa (the capital of the Yemenite Arab Republic) the mean February temperature is only 12°C, and that of August is 25°C.

Similar features of temperature distribution are characteristic of Eritrea, situated on the west coast of the Red Sea between 12°40' and 18° NL and 36°30' and 43° E. long. On the coast itself the climate is hot and dry. In Massawa, for example, the mean temperature for July is 35.0°C, and for January, 26.0°C. Small amounts of precipitation usually fall during the winter, totaling up to 200–250 mm annually. In the internal, mountainous regions of Eritrea, the climate is more humid and cool. According to the data of the Meteorological Station of Adi-Ugri situated at a height of 2022 m, the air temperature during the summer months does not exceed 21.5°C. In the mountainous part of the country the annual amount of precipitation reaches 500–1000 mm in some places, and is concentrated mainly during the summer months.

Winter air temperature

The distribution of the mean temperature over extensive territories, as well as over the entire terrestrial globe, is conveniently represented by isotherm maps. Isotherms are lines which connect points having the same temperature values. The temperature is reduced to one level, usually to sea level, so that data at various points can be compared*. This is done because the air temperature drops, as a rule, with increasing height, and thus the temperature high in the mountains is lower than in nearby valleys. The reduction to one level, in particular to sea level, is done on the basis of the well-known fact that for each 100 m of ascent the temperature drops on the average by 0.5° –0.6°C.

Figure 18 is a map of the mean temperature for January, reduced to sea level. The nonuniformity in the temperature distribution over the terrestrial globe is shown by the odd shapes of the isotherms, particularly on the boundaries between sea and land. The isotherms are not parallel, as they should be if they were in accordance with the amount of solar energy received.

Following the zero isotherm, we see that to the north of the Pacific Ocean it is situated near 60° NL and over North America it passes somewhat to the south of 40° NL. In other words the zero isotherm is displaced

* The observation data refer to a height of 2m, since instruments at Meteorological Stations are mounted at this height.

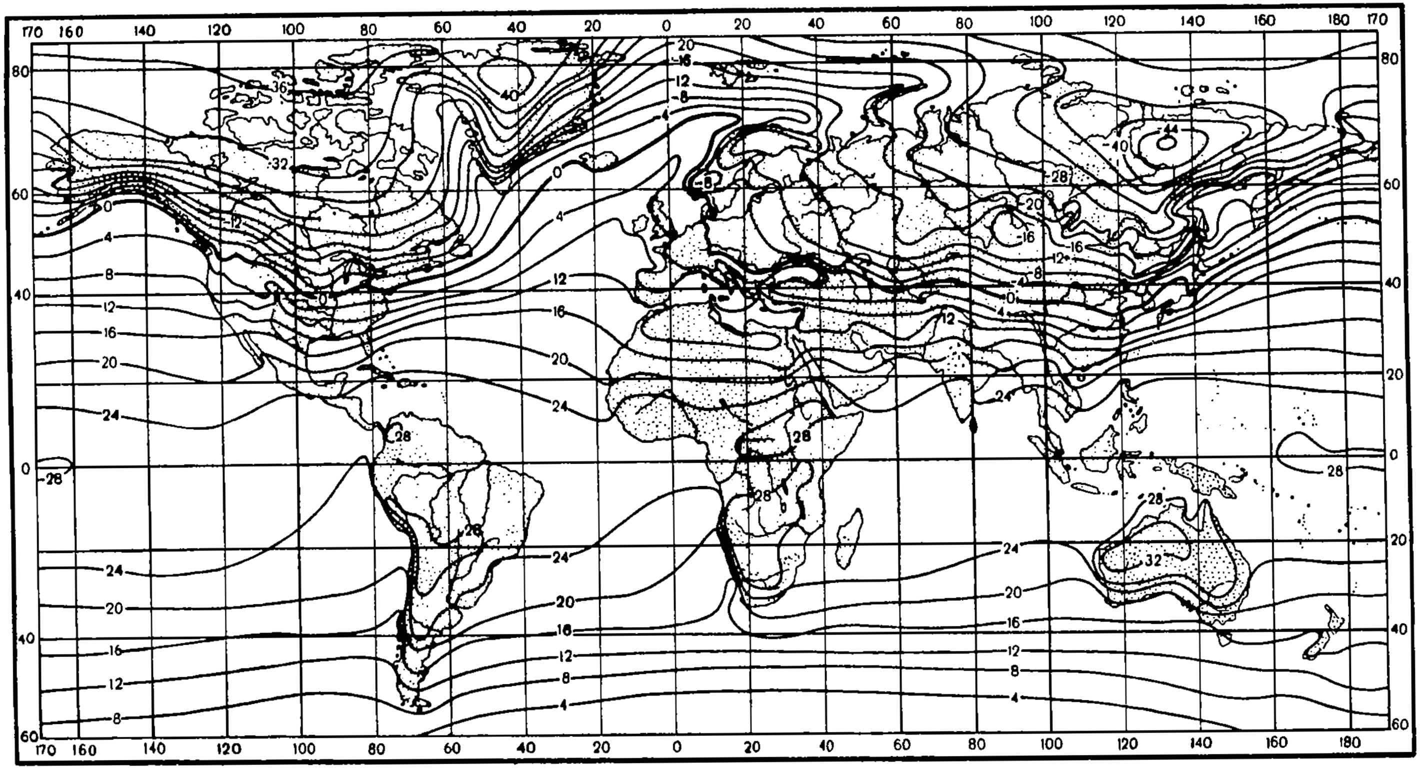

FIGURE 18. Isotherms for January

by 20° along the meridian; this amounts to about 2200 km. Following the shore of America, the isotherm reaches the 70° latitude to the north of the Norwegian sea. Going around the Scandinavian peninsula, it drops through Central Europe to the Danube basin and, going east, goes through the territory of the Chinese People's Republic to the south of 34° NL.

The mean air temperature is thus generally the same at the extreme north of the Atlantic, at a latitude of about 70°, and in Central China at a latitude of 34°, although the distance between these latitudes is approximately 4000 km. The remaining isotherms at middle latitudes have the same configuration.

It can be easily shown that this is a result of the difference in the physical properties of the underlying surface of the Earth, i.e., of the warm ocean water and of the cold continents. This factor determines the concentration of the isotherms at the shores of the continents. Furthermore, it can be seen that the concavity of the isotherms on the continents of the northern hemisphere is directed toward low latitudes, and in the southern hemisphere toward high latitudes. This happens because January is a winter month in the northern hemisphere and a summer month in the southern hemisphere. Therefore in the northern hemisphere the continents are highly cooled, and the ocean water, still possessing sufficient heat reserves, heats the air masses flowing over them. Meanwhile, in the southern hemisphere during the southern summer the continents are heated more than the oceans.

Another peculiarity of the temperature distribution over the terrestrial globe in January is the higher isotherm density in the northern hemisphere as compared with the southern hemisphere. This happens because during the winter in the northern hemisphere an intensive cooling of the continents and the air layers adjacent to them takes place at high and middle latitudes, whereas to the south of 40° NL, heating prevails. Only in the equatorial zone the inflow of solar energy, and consequently the temperature does not undergo substantial variations during the year. Thus, large temperature differences between the tropics and the north are created. In the southern hemisphere in the middle of the southern summer, the balance of solar energy is positive everywhere and the interlatitudinal temperature difference decreases.

On the continents, cold and hot regions are observed in both hemispheres. In January the cold focuses are situated to the north of the Equator, and heat focuses – to the south. The largest cold focus, edged by the -48°C isotherm, is situated in the north of Yakutia, and a smaller one exists over Greenland (it is outlined by the -42° C isotherm). Smaller cold focuses are observed over Scandinavia, Asia Minor, and other regions of the Eurasian continent. There are heat focuses in the southern hemisphere, over South America, Africa, and Australia.

In the example of the path taken by the zero isotherm, we already saw that at similar latitudes the mean January temperature varies sharply.

The differences between the observed mean monthly and latitude-averaged temperatures characterize the influence of the continents and of the oceans on the temperature regime of the air near the Earth's surface. These data show that in the northern hemisphere in January, both on the continent and on the oceans, the deviations of the temperatures from the latitude-averaged values are large.

The largest positive deviations are observed over the oceans, and the largest negative deviations — over the eastern regions of the continent. Over the Norwegian Sea, between 60° and 70° NL, the mean January air temperature is 26°C higher than the latitude-averaged temperature. To the north of the Pacific Ocean the largest positive deviations are observed over the Gulf of Alaska but they do not exceed 13°–14°C. In western Europe, due to the action of the Atlantic and of the warm Gulf Stream, the positive temperature deviations sometimes reach 10°–12°C. For this reason the mean temperature in Leningrad in January exceeds the latitude-averaged temperature by 8°C, in Murmansk by 20°C, on the Spitsbergen islands by 16°–20°C, and in Arkhangelsk by 9°C.

The negative temperature deviations from the latitude averaged values reach 4°C in Sverdlovsk, 8°C in Novosibirsk, 16°C in Chita, and even 24°C in Verkhonyansk and Oimyakon. From the position of the line of the zero temperature deviation we cannot conclude that the effect of the Atlantic Ocean is confined to eastern Europe. Investigations show that in the northern regions of western and eastern Siberia, relatively warm and humid air masses from the Atlantic often arrive through the Soviet Sector of the Arctic. Heat transfers from the direction of the Pacific Ocean are not infrequent on northeast Asia.

In North America the maximum negative temperature deviations from the latitude-average are observed in Canada between 50° and 60° NL. The influence of the Pacific Ocean on the temperature regime of the air at the surface of the Earth is confined mainly to the coastal zone, since the meridionally lying high Cordillieras and Rocky mountains obstruct the way of warm and humid air masses coming from the direction of the Pacific Ocean to the continent.

As one approaches low latitudes, the temperature deviations decrease, and in the equatorial zone they do not exceed 2°–3°C.

During the summer the situation is quite the reverse, over the oceans negative deviations and over the continents —mainly positive deviations are observed, and their absolute values are less than a half those of the winter.

In the southern hemisphere, owing to the predominance of ocean surface and to the smallness of the continents, the temperature deviations from the latitude-average do not exceed 4°–8°C during the summer. In both hemispheres regions of negative temperature deviations exist in the zone of cold currents during the summer, e.g., the Canaris and Californian regions of the northern hemisphere, and the Peruvian and Benguela regions in the southern hemisphere. The influence of these cold currents on the mean air temperature is clearly displayed also on the isotherm maps of January (Figure 18) and of July (Figure 19). The influence of the Californian current is particularly large, producing considerable temperature differences between the land and the ocean during the summer.

Summer air temperature

The air temperature distribution at the surface of the Earth undergoes significant variations from winter to summer. During the month of July the continents in the northern hemisphere are heated more than the oceans.

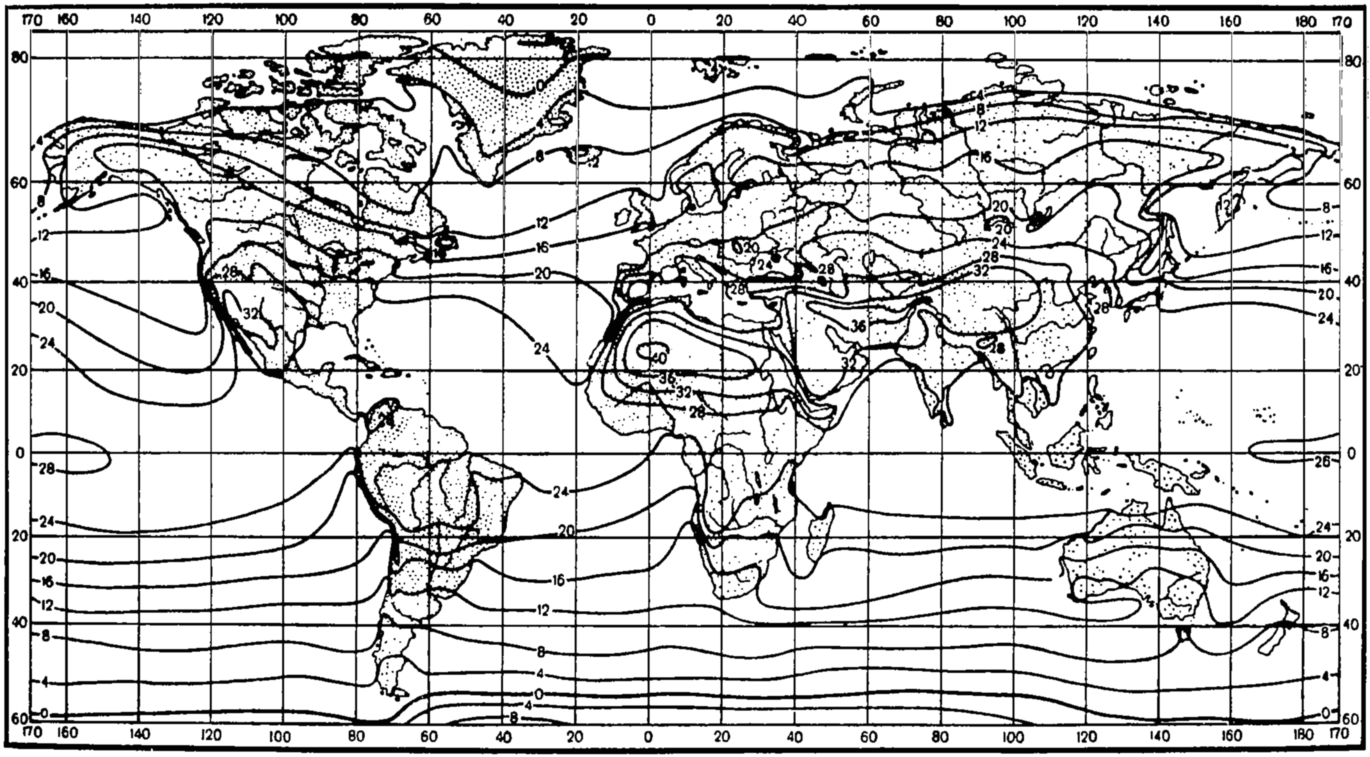

FIGURE 19. Isotherms for July

Over North America, Africa, and South Asia distinct heat regions exist (Figure 19). The large horizontal temperature variations between the continents and the oceans, observed during the winter over Alaska, the south of Greenland, and the east of Asia, disappear during the summer. They reappear in turn on the west coast of North America and on the west coast of North Africa. As was indicated above, these variations are created because the heating of the land interacts with the cold ocean currents.

In the southern hemisphere the annual isotherm configuration does not vary appreciably; the exceptions are those continents where isolated heat regions disappear during the winter. It is natural, however, that in various parts of the terrestrial globe the absolute values of the air temperature vary differently from summer to winter. Whereas in the equatorial zone the differences between the mean temperatures of the cold and warm months usually do not exceed 2°-3°C, over Yakutia these differences reach 60°-65°C.

The absolute values of the temperature differences between the winter and summer months reach even larger values on land. The lowest temperatures in the northern hemisphere are observed at Verkhoyansk and Oimyakon (Yakutia). These points are situated at different latitudes and at different heights above sea level. But the important factor in the physicogeographic conditions is the air stagnation caused by the rather unique topographic conditions and the character of the atmospheric circulation. Verkhoyansk and Oimyakon are situated at the bottom of craters surrounded by mountains. The stagnated air in the craters is strongly cooled during the winter and considerably heated during the summer. Situated far from the warm water surfaces of the Atlantic and protected from the Pacific Ocean by high mountains, Yakutia is subjected to very little outside influence, particularly during the winter. Since the air circulation there is not particularly intense, the lower air layers are cooled during the winter months and heated during the summer. The lowest temperature observed in Verkhoyansk is -67.8°C, and in Oimyakon, -70°C. The highest temperature recorded in Verkhoyansk is 33.7°C. Thus, the difference in temperature between the absolute minimum and the absolute maximum exceeds 100°C there.

In other middle-latitude regions far from seas and oceans the difference between the absolute maximum temperature and the absolute minimum temperature exceeds 50°–60°C annually. Thus in Moscow, for example, temperatures of -25°C and +25°C are recorded nearly every year. Even during one month (January) temperatures of -42.2°C and +4.9°C were recorded in Moscow – a difference of 47.1°C. The extremal temperatures in Moscow during the last 80 years were -42.2°C (January) and +36.8°C (July) – a difference of 79.0°C.

Even the mean temperature of the same month undergoes considerable variations from year to year, particularly the temperature during the winter months. In Moscow the mean January temperature for many years was -9.7°C. For 1893 it was -20.5°C, and for 1925, -3.3°C. During December, 1960 the monthly mean temperature was higher than the many-year mean by 8.2°C (+0.2°C instead of the average -8.0°C). Warm weather was maintained during the whole month and continued during the first ten days of January, 1961. This was the longest period of heat anomaly during the last 80 years.

Heat and cold anomalies usually extend over large regions and depend on the character of the atmospheric circulation on at least a hemispheric scale. In particular, the warm, almost snowless weather of December, 1960 in the European territory of the USSR was caused by the prevailance of zonal circulation, associated with the motion of cyclones over northern Europe and western Siberia. Air masses moved in from the Atlantic after being heated over the warm surface of the ocean. On the other hand, unusually low temperatures during several days and months of the winter were observed when air masses moved in from the Arctic and from Siberia.

Due to the general heating of the underlying surface, the temperature anomalies during the summer do not reach such high values as those of the winter. In Moscow, the highest mean temperature of July over the last 80 years, equal to +23.3°C, was observed in 1938, the lowest, 14.6°C, in 1935. The absolute maximum was +36.5°C and the absolute minimum was +1.3°C, yielding a difference of only 35.2°C, instead of 47.1°C in January.

Air temperature in the Arctic and the Antarctic

Data on atmospheric processes in the Arctic and in the Antarctic have only been collected during the last 20-30 years, although these inaccessible regions of the Earth have attracted the attention of investigators for a long time.

During the 1920's, the need for utilizing the northern sea route posed a series of new problems for Soviet scientists. They began a systematic study of the Central Arctic during the thirties, since without a thorough study of the ice-floe regime and of the hydrometeorological processes that create it, it would have been impossible to determine navigation changes.

In 1932 Soviet scientists in the Arctic took an active part in the 2nd International Arctic Year program. Organized by the Central Administration of the northern sea route, a wide network of scientific stations along the northern coast of the Soviet Union were set up. In 1937 the first investigators of the Central Arctic, I.D. Papanin, E.K. Fedorov, P.P. Shirshov, and E.T. Krenkel', landed near the North Pole. After the ice floe on which the "Severnyi Polyus" Station was situated had drifted for a long time it was carried away from the Central Arctic to the Greenland Sea. In February, 1938 the staff of Papanin's expedition were taken off the ice floe not far from the western shores of Greenland, at the latitude 70°54'.

The first valuable results of investigations of the Central Arctic were given by the expeditions on the drifting scientific station "Severnyi Polyus" (1937-1938) and on the icebreaker "Georgii Sedov" (1937-1940). The observational data obtained during these expeditions widened our knowledge of the ice-floe regime, the meteorological processes, and magnetic phenomena. They helped in solving many problems related to weather forecasting and the conditions of the ice in the Arctic. The observational data of the drifting station "Severnyi Polyus" disproved the assumption that a high atmospheric pressure with a relatively calm anticyclonic weather and constant low temperatures are characteristic of the central part of the Arctic. The observations showed that cyclones appear frequently in the Arctic, bringing with them overcast conditions, snowstorms and storms which are accompanied by short rises in temperature.

These data also changed the prevailing ideas on the atmospheric circulation in the Arctic. It was established that in the Central Arctic masses of warm air penetrate not only from the Atlantic Ocean but also from the Pacific Ocean through Chukot Peninsula and Alaska. Aerological observations showed that warm air masses, even in the conditions of the Polar Night, cool down quite slowly, propagating over the surface layer of cold air.

On 2 April, 1950 the scientific station "Severnyi Polyus-2" (SP-2) was set up on drifting ice to the north of Wrangel Island; a year later, on 9 April, 1951, it was dismantled. New scientific stations on drifting ice were set up during April, 1954; the SP-3 station on 9 April, the SP-4 station on 6 April. During one year, both stations drifted a distance of over 2000 km. From that time on, drifting stations were being organized systematically.

The scientific staff of these stations collected, and still collect, valuable information concerning various aspects of the meteorological processes and the ice-floe regime of the central Arctic.

Recently it was established that sharp rises in the temperature of the Arctic during the winter are characteristic not only of the troposphere. Investigations have shown that in the stratosphere, at heights of 20–30 km, the temperature rises considerably for short periods of every year. In some individual cases the air temperature increased during the Polar Night in the course of 5–10 days by 20°–30°C and more.

Discoveries were made not only in meteorology, but also in sea-hydrology, magnetology, hydrobiology, and other fields.

The work of the courageous investigators of the Arctic is carried out under difficult conditions. The Arctic can spring unpleasant surprises at any moment. Strong frosts can be replaced by short-lived temperature rises that bring precipitation, snowstorms, ice-jamming, etc.

The Antarctic was discovered by F.F. Bellingshausen and M.P. Lazarev in the first quarter of the 19th century. However, until now its nature was not adequately known.

Investigations carried out during the International Geophysical Year (1957–1959) furnished many more data than all the investigations made during many tens of years after the Antarctic was discovered.

At stations situated all over the Antarctic, large and multinational groups of scientists conduct investigations of geophysical processes and phenomena under the conditions of the most severe climate of the world. They study the weather, climate, circulation of the atmosphere, and the aurora polaris. Radiosondes, radar pilots and meteorological rockets sound the troposphere and the stratosphere. Glaciologists pierce the two-three-kilometer thickness of the ice plateau of the continent. Geodesists determine its height above sea level. The regime of the sea currents, the life in the ocean, and other topics are also studied.

Situated at high latitudes of the southern hemisphere, the sixth continent is always covered by ice whose thickness reaches 2000 m. According to calculations, if the ice of the entire Antarctic melted, the level of the oceans would rise by 50 m and the water would flood many coastal areas.

Recent calculations concerning the balance of the Antarctic ice showed that about 130 mm of precipitation fall on the average per year; this is equivalent to 1830 km^3 of water over an area of 14.1 million km^2. The annual loss of ice by the breaking up of the icebergs amounts to only 730 km^3. Estimating by these approximate calculations, the ice armor of the Antarctic grows annually by 1100 km^3.*

The lowest air temperatures on Earth were recorded on the high ice plateau of the Antarctic. In the interior area of the Antarctic, at about 3000 m above sea level, the mean annual temperature is -36.5°C. This is the lowest mean air temperature anywhere on Earth. During the winter the air temperature often drops to -70° or -80°C, and sometimes even lower. Thus, for example, on 10 August, 1958, at the Sovietskaya station, a temperature of -86.7°C was recorded; on 25 August the temperature at the station Vostok fell to -87.4°C; at the Yuzhnyi Polyus [South Pole] station the lowest temperature of -74.7°C was recorded before sunrise. At the station Vostok on 24 August, 1960, the lowest temperature on the terrestrial globe, -88.3°C, was recorded.

At the station Vostok, situated at approximately 78° SL, a lower temperature (-88.3°C) was recorded than at the South Pole itself (-74.4°C). This is because the Vostok station is situated on a plateau where in windless weather the air cools down strongly under the conditions of the Polar Night, whereas the Yuzhnyi Polyus station is situated closer to intense cyclonic activity in the region of the Ross Sea.

Very low temperatures were also observed in the central Antarctic during the summer months.

Investigations show that over the central Antarctic air temperatures of between -45° and -50°C are most often observed during the winter, and that this varies only slightly when the temperatures at the surface of the Earth are very low. Hence it follows that the above-mentioned low temperatures are characteristic of the surface air layer. The temperature increases rapidly with height, but only up to a level of 4–5 km; there it starts to decrease with height (Figure 20).

In the Arctic the mean air temperature is higher than in the Antarctic all year around. During the winter near the Pole it is higher by 7°–8°C, at 80° latitude – by 13°C, and at 70° latitude – by 10°C. During the summer these differences increase (see Table 11).

This increase is due to the difference in the character of the atmospheric circulation and of the ocean currents in the Arctic and in the Antarctic. Owing to the positions of the continents and of the oceans in the northern hemisphere, the western transfer in the troposphere is very often upset by intensified interlatitudinal large-scale exchange. Consequently the temperature in the Arctic often rises during the summer to 5°–10°C, and in individual cases – up to 20°C and more. During the winter thaws are observed. The absence of land in the Arctic region contributes to the development of ocean currents, owing to which large, continuously moving icebergs drift towards the Greenland Sea and are carried to the northern part of the Atlantic Ocean.

The southern borders of Africa and Australia do not extend to beyond 40° latitude, nor the southern tip of South America beyond 55° latitude. There is however a huge continent with a thick ice armor in the center of the Antarctic. The interlatitudinal air exchange in the Antarctic is therefore weaker than in the Arctic. As can be seen from the maps, the winter and summer isotherms span the southern hemisphere almost along the parallel.

* See the paper: Shcherbakov, D. Tainy shestogo kontinenta (Secrets of the Sixth Continent). — "Pravda", 27 June, 1961.

It does not follow, however, that the atmospheric processes are less intense in the southern hemisphere. On the contrary, the large interlatitudinal temperature and pressure differences give rise to deep cyclones with large horizontal pressure gradients and gale winds between middle and high latitudes. Even the mean velocities reach 10-15 m/sec, and the frequency of gales is higher than 50%.

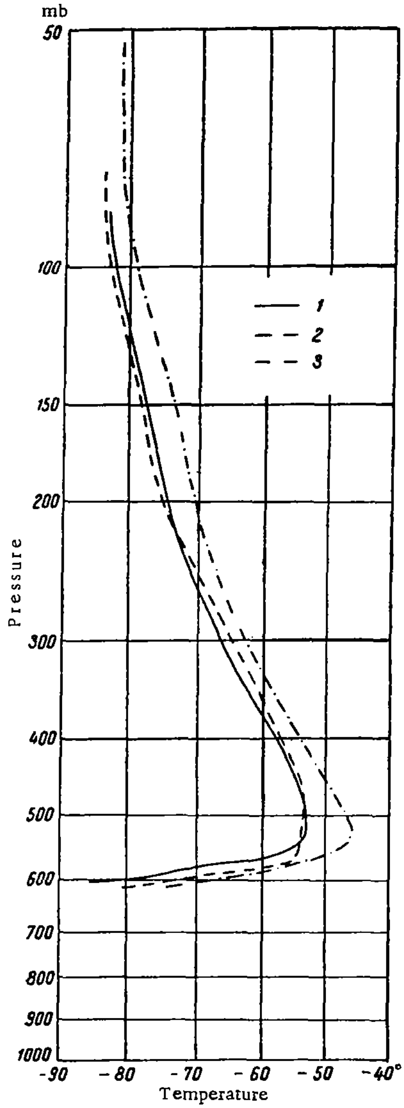

FIGURE 20. Variation of temperature with height over the central Antarctic during the winter for low temperature at ground level

1 – Vostok station, 25 Aug., 1958; 2 – Vostok station, 23 Aug., 1960; 3 – Sovietskaya station, 11 Aug., 1958.

The variation curves of the monthly mean temperature with height in the Arctic and in the Antarctic (Figure 21) indicate that during the winter in the central Antarctic the mean temperature at all heights is lower than in the central Arctic. During the summer the same picture is observed in the troposphere, but in the stratosphere, above 16 km, the temperature difference between the Poles disappears. In addition, the temperature difference between summer and winter in the South Pole is appreciably larger than in the region of the North Pole.

As was already mentioned, the difference between the temperature distributions with height in the Arctic and in the Antarctic are caused mainly by the circulation peculiarities in the northern and in the southern hemispheres. As a result of the large-scale interlatitudinal air exchange at high latitudes of the northern hemisphere during the winter, relatively warm air arrives in the central Arctic from medium latitudes more often than in the internal regions of the Antarctic; there the interlatitudinal air exchange is of a smaller scale.

The cyclonic activity at the shores of the Antarctic is associated not only with strong winds, but also, mainly on the periphery of the continent, with precipitation, where the annual amount is several hundreds of millimeters.

As one approaches the South Pole the influence of the cyclones weakens and the amount of precipitation decreases from north to south. In the central part of the plateau the amount of precipitation is small. Yet ice still accumulates in the Antarctic and slides under its own weight to the shores, falls into the ocean, and forms a multitude of icebergs.

According to approximate data, the thickness of the ice cover of the Antarctic varies from some hundreds to 4000 m. By means of seismic methods it has been established that in places the ice sheet of the continent penetrates below sea level. This served as a basis for the assumption that the Antarctic is not one continent, but is made up of islands. However, data of special Soviet expeditions in the Antarctic, which traveled from Mirnyi to the South Pole and to the zone of relative inaccessibility, justify the alternative idea that the Antarctic is nevertheless a continent and not an archipelago.

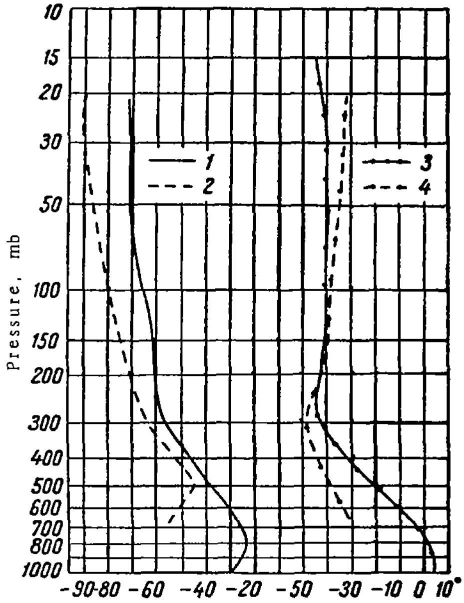

FIGURE 21. The variations of the monthly mean temperature with height over the central Antarctic and the central Arctic during winter and summer

1 – "Severnyi Polyus-7" station, January; 2 – Amundsen-Scott station, July; 3 – "Severnyi Polyus-7" station, July; 4 – Amundsen-Scott station, January.

Very low temperatures are also observed on the ice plateau of Greenland. Situated at high latitudes of the northern hemisphere, the largest island of the world is covered to a large extent by eternal ice with a mean thickness of 1500–2000 m. The lowest temperatures are observed on the ice plateau in the middle part of the island. There, at a height of about 3000 m above sea level, the annual mean air temperature is about 30°C below zero and varies between cold and warm months from -47°C to -11°C. This is due to the considerable height above sea level, since in the Arctic at the same latitudes the monthly mean air temperatures are higher. Very low temperatures are also observed during the summer months. Severe frosts and a small annual amount of precipitation (less than 100 mm) is also characteristic of the northern end of Greenland (to the north of the 80th parallel).

At the same time on the southern coast of the island, situated at the latitude of Leningrad and Arkhangelsk, only moderate frosts are observed and an abundant amount of precipitation falls. The mean air temperature of the coldest month reaches only 6°-10°C below zero, and the annual amount of precipitation exceeds 1000 mm in places. The influence of the cold central part of the island is weakened there to some extent by the "warm breathing" of the Atlantic. The cyclonic activity, characteristic of the southern part of Greenland the year around, is associated not only with the fall of excessive amounts of precipitation, but also with strong gale winds. A continuous transfer of ice from the Arctic takes place through the Greenland Sea.

The role of sea currents in heat transfer

Sea currents play an important role in the formation of the air temperature field over the terrestrial globe. We have already seen the size of positive deviations of the air temperature from the latitude-averaged value on the north of the Atlantic and Pacific oceans during the winter. This is due to the capacity of the ocean water to accumulate the heat received during the summer and to lose it during the winter by heating the air. However, the high temperatures of the surface water and air are caused also by the transfer of heat by sea currents from low to high latitudes, as well as from high to low latitudes.

Water in the oceans and seas is in continuous motion. Sea currents are caused mainly by the action of the wind and by water-pressure differences arising at the same levels. Sea currents are also affected by the surface and internal friction forces and by the Coriolis force. By the action of the latter, the wind-generated current is deflected to the right of the wind direction through an angle of 45°. The higher the current velocity, the greater the influence of the deflecting force of the Earth's rotation. This force has no effect on sea currents at the Equator. In the southern hemisphere the currents are deflected to the left.

In shallow water basins, wind currents, or, as they are usually called, drift current, are only slightly deflected from the direction of the wind. In oceans and deep seas, on the contrary, the surface and internal friction forces and the Coriolis force complicate the structure of the sea currents. Theoretical calculations show that the velocity of a drift current decreases considerably with depth. Drift currents usually entrain surface water layers several tens of meters thick. With depth, the current is deflected to the right up to the reverse direction, a process associated with a decrease in velocity. However, as shown by calculations, the decrease in the velocity is so rapid that the total flux of the drift current is directed to the right of the wind direction at right angles to it.

By simultaneously examining sea-current and wind maps it is possible to discover the common features in the geographic distribution of the wind and of the currents; these point to a close interaction between the atmosphere and the hydrosphere of the Earth. In fact, the continents and the oceans affecting the thermal regime of the atmosphere upset the western transfer; this transfer is due to the general inequality of the solar energy

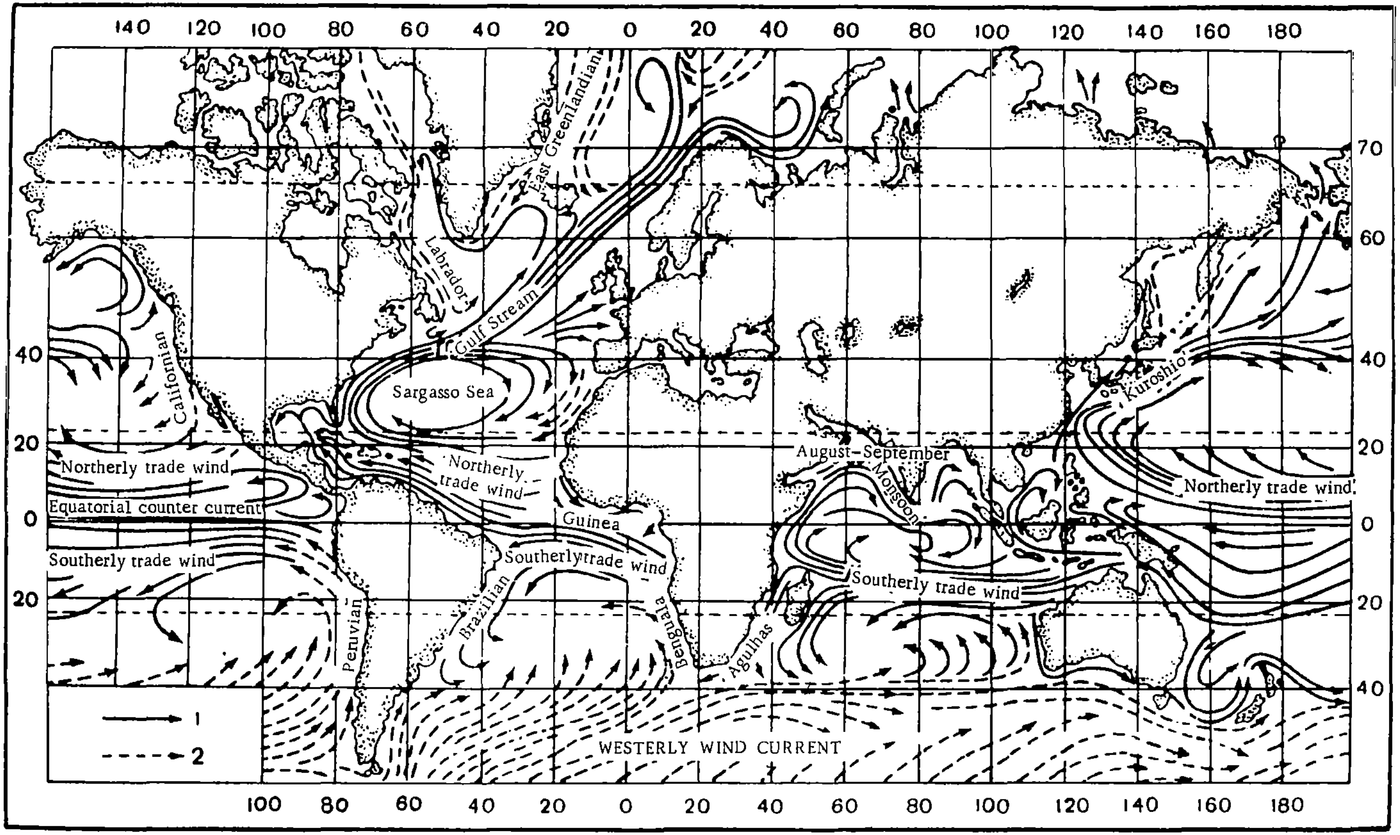

FIGURE 22. Map of sea currents, January

1 – warm currents; 2 – cold currents.

inflow at the Equator and at the Poles. Consequently, air currents in some regions are diverse, and in other regions – have a seasonal constancy.

Under the constant action of the wind, the water in the oceans is brought into motion in a direction determined by the action of a number of forces.

A certain influence is exerted on the direction and velocity of the ocean water by the position and form of the continents.

Figure 22 shows the circulation of the oceans during January. One notices the powerful currents in the Atlantic and Pacific oceans, and in the region of the trade winds between 0 and 20° latitude. At these latitudes the air motion is characterized by seasonal constancy; the northeasterly trade wind in the northern hemisphere and the southeasterly trade wind in the southern hemisphere. Under the action of the trade winds, the rising sea currents turn to the west. As can be seen from the map, low velocities are observed near the continents which constitute a natural obstacle. Thus, for example, the water of the Atlantic, flowing from the equatorial zone to the west and encountering on its way the northern coast of South America, reduces its velocity and flows along the continent. Similar decrease in the current velocity can be observed at the southeastern coasts of northern America and of Asia.

Under the action of the trade winds, the water of the equatorial zone of the Atlantic, turning to the region of the Caribbean Sea and of the Gulf of Mexico, form to the east of Florida the most powerful warm current in the world – the Gulf Stream. The Gulf Stream waters pass through the entire northern half of the Atlantic Ocean, reach the Norwegian and Barents seas, and the solar heat contained in them, which was accumulated in the equatorial zone and in the tropics, warms not only the northern seas during the winter, but also large areas of Europe and Asia.

Similarly, another warm current, called the Kuroshio, arises at the eastern coasts of the Pacific Ocean. From the equatorial zone of the Pacific Ocean, also under the action of the northeasterly trade wind, the water turns to the west. However, the Asian continent and the numerous islands situated in the way of this current diversify it. One part of the current is directed to the Celebes Sea, another – to the South China Sea, and only the remainder of this powerful northern equatorial trade-wind current reaches the China Sea and the Japanese islands. The Kuroshio is therefore considerably inferior to the Gulf Stream. In this way also cold counter currents arise, as for example, the east Greenland, the Labrador and the Canaris cold currents in the Atlantic, the Kamchatka, Californian, and Peruvian cold currents in the Pacific Ocean and others.

Warm and cold currents in the oceans carry a huge amount of heat.

Calculations of the amount of heat received or lost by the surface ocean water due to sea currents, performed by M.I. Budyko, T.G. Beryland, N.I. Zubenok and others (Central Geophysical Observatory), showed that the Gulf Stream alone transfers annually from the equatorial zone to the North Atlantic an amount of heat equal to 80-100 kcal/cm^2 (Figure 23).

In Figure 23 the isolines represent the distribution of the heat received or lost by the oceans due to its transfer by sea currents. It can be easily seen that the heat transfer by the Kuroshio current on the west of the Pacific Ocean is considerably weaker than the heat transfer of the Gulf Stream. Near the Japanese islands, the amount of transferred heat reaches only 20–30 kcal/cm^2·year.

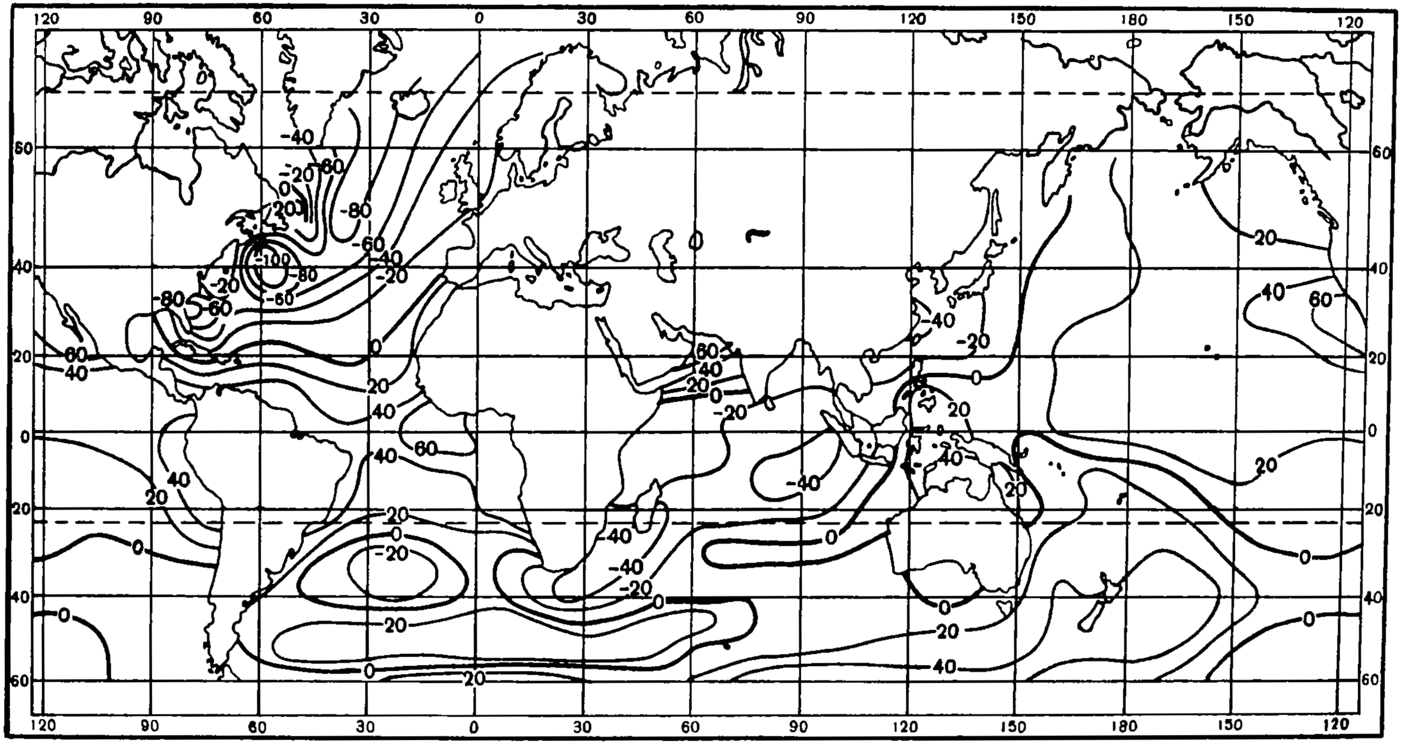

FIGURE 23. Amount of heat received or lost by the ocean surface due to the action of sea currents, kcal/cm^2.year

To get an idea of this amount of heat transported to the north, we note that in the northern hemisphere between 40° and 60° NL, from 80 to 120 $kcal/cm^2$ of the total solar radiation fall on the middle part of the oceans per year.

In the cold current zone, the ocean loses a considerable amount of heat. Thus, for example, between 20° and 40° NL at the Californian coast the ocean loses up to 60 $kcal/cm^2$. year, and the amount of heat received by the ocean in the form of incoming total radiation is equal to only 100 $kcal/cm^2$. year. This is why in this part of the Pacific Ocean the isotherms on the January and July maps (see Figures 19 and 20) are bent towards low latitudes and the difference between the air temperatures over the cold water of the ocean and the heated land is quite large. The transfer of heat by the ocean in the zone of the Canaris cold current at the western coast of North Africa is somewhat weaker, but this too is reflected in the climatic conditions of these regions. The water surface exerts a decisive influence on the air temperature. Similar to the smooth variation of the temperature of the surface water, the temperature of the air over a water surface also has a smooth annual variation. The temperature difference between winter and summer months over oceans and seas is considerably smaller than over land. For example, in the region of the Azores Islands, as well as at 60° latitude (to the south of Iceland) the difference between the mean air temperatures of January and July does not exceed 7°–8°C. For the same period over land in the region of Moscow, this difference reaches 27°C, and in Sverdlovsk 34°C.

The extent of the influence exerted by the seas on the temperature of the surface air layer can be shown by the example of the Black and Caspian seas. Both these seas protect the Caucasian coast, particularly Transcaucasia from strong frosts. The path of the cold air masses from the north is blocked by the Caucasus Range, whose mean height is 3–4 km. When cold air masses penetrate through the Black or Caspian seas from the west or from the east, they rapidly change their properties while moving over the water surface and become warmer and more humid in their lower layers. However, when moving rapidly over such a small water basin as the Black Sea, the air has no time to rapidly heat up and its penetration into the subtropics sometimes causes the ruin of heat-loving cultures. Thus, for example, during the winter of 1949–50 the sharp cooling off on the Black Sea coast of the Caucasus, related to the penetration of cold air masses from the north through the European territory of the USSR, ruined the citrus plantations.

Smaller water basins also exert an influence on the air temperature; however, their radius of action is very limited. For example, the influence of the deep-water Lake Baikal on the air temperature both during the winter and summer is felt within a radius of 100–200 km. The influence of Lake Sevan, whose area is 1400 km^2, extends to only 10–15 km during the winter, and during the summer, due to the small difference between the water and land temperatures, is even smaller.

MOTION OF THE AIR

The reason for air motion

The air surrounding us is in continuous motion. We feel this as a wind.

The inflow of solar energy, the nonuniformity of the underlying surface, and the rotation of the Earth about its axis are the main factors causing air currents. How is the arriving solar energy converted into the kinetic energy of air masses? This is one of the basic problems of meteorology. Its theoretical solution would enable us to find the pattern of large-scale atmospheric processes and thereby to forecast them, thus creating a scientifically based method of long-range weather forecasting. To solve this complicated problem, we use the equations of thermodynamics and hydrodynamics in a form applicable to the conditions of the atmosphere.

The problem is a very difficult one. During the last 20 years considerable progress has been made. A distribution, close to the real one, of the temperature and zonal wind on the terrestrial globe was obtained by calculations.

Let us return, however, to the question of the reasons for air motion.

Under the action of the above-mentioned factors a nonuniform air pressure distribution arises on Earth. As a result of the atmospheric circulation, the pressure field, and consequently, also the air currents, undergo continuous variations.

Air flows from high-pressure to low-pressure regions, and continues until the pressure difference disappears. The vector characterizing the degree of variation of the atmospheric pressure in space is called the pressure gradient.

The pressure gradient is equal to the pressure variation per unit distance. For this unit 1° of meridian (111.1 km) is taken. The larger the pressure gradient, the higher the wind velocity. The pressure gradient is directed along the normal from high to low pressure.

In the Soviet Union the wind velocity is determined in meters per second (m/sec) and in kilometers per hour (km/hr). At sea the wind force is measured in numbers on the twelve-unit [Beaufort] scale. In a number of other countries the wind velocity is measured in miles per hour. To convert the wind velocity from m/sec into km/hr and from Beaufort numbers to m/sec Tables 17 and 18 can be used.

The large- and small-scale nonuniformity in the pressure distribution is determined by the action of forces arising when the thermal solar energy is transformed into kinetic energy of the air.

The heating and cooling of the air in the lower layers of the atmosphere is caused by the Earth's surface. Upon heating, the air expands and ascends upward, colder air dropping to replace it. The degree of heating of

TABLE 17

Conversion of wind velocities from m/sec to km/hr

m/sec	1	2	3	4	5	6	8	10	12	15	20	25	40	50
km/hr	3.6	7.2	10.8	14.4	18.0	21.6	28.8	36.0	43.2	54.0	72.0	90.0	144.0	180.0

Conversion of wind velocities from km/hr to m/sec

km/hr	5	10	15	20	25	30	40	50	60	80	100	150	200
m/sec	1.4	2.8	4.2	5.6	7.0	8.3	11.1	12.9	16.6	22.2	25.8	23.0	56.0

TABLE 18

Conversion of wind velocities from Beaufort numbers

Beaufort numbers	0	1	2	3	4	5	6	7	8	9	10	11	12
Velocity, m/sec	0—0.5	0.6—1.7	1.8—3.3	3.4—5.2	5.3—7.4	7.5 —9.8	9.9—12.4	12.5—15.2	15.3—18.2	18.3—21.5	21.6—25.1	25.2—29.0	Over 29.0

the surface of the Earth, and consequently also of the air, is different not only at points that are far apart, but also at those near to each other. The process of heating and of the replacement of warm by cold air masses thus takes place continuously and everywhere.

Everyone of us has observed the setting up of air motion: when the front door is opened during the winter, cold air rushes into the warm room and air escapes outside. It is obvious that the reason for this phenomenon is the difference between the air temperature in the warmed room and that of the outside air.

The conditions leading to the setting up of air currents can be described in an elementary way as follows.

Suppose that in neighboring regions a and b over water and land, respectively, the pressure at the surface is P_0, and at some height, P_h. Let the mean temperature of the air layer between these surfaces be T_2 over the water and T_1 over the land. Schematically, the distribution of the air masses in the vertical plane over these surfaces can be represented as shown in Figure 24.

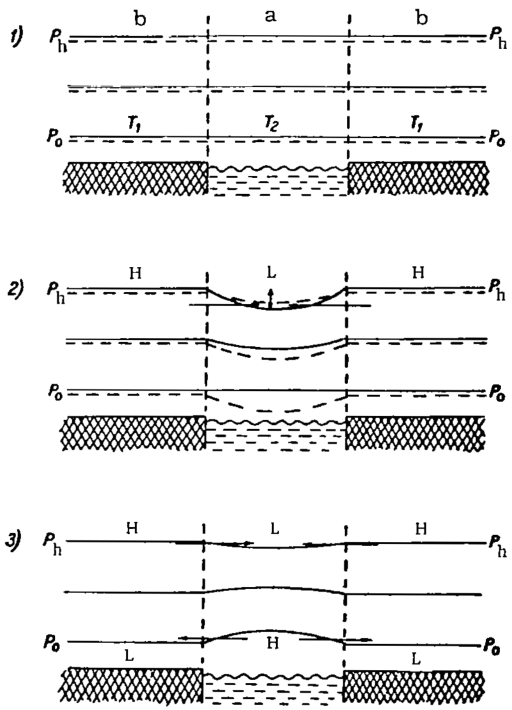

FIGURE 24. Air motion arising due to a temperature difference

1 – equal air temperature and pressure over the neighboring regions a and b; 2 – air temperature over the water surface lower than over the land; 3 – structure of the isobaric surfaces and of the wind (shown by arrows) at the surface of the Earth and at different heights.

Suppose that the indicated neighboring regions are initially in thermal equilibrium, i.e., $T_1 = T_2$, and that the air pressure both at the surface of the Earth and at various heights is uniform. This means that the isobaric surfaces, i.e., the surfaces with equal pressure, coincide with the horizontal surfaces; this is shown in the first diagram of Figure 24. In this case the air masses are at rest. To upset the equilibrium let us assume that the air over the region a (over the water surface) begins to cool down. After some time its mean temperature T_2 is lower than the air temperature T_1 in the neighboring regions b (over land). The cooling of the air is depicted by the bending of the broken lines, i.e., of the temperature isolines, downward. It is natural that initially the greatest rate of cooling is observed at the surface layer; with increasing height it decreases. Owing to the cooling of the air over the region a its density increases, and the isobaric surface P_h at the upper level descends by an amount shown by the arrow in Diagram 2 (Figure 24).

This can be explained as follows: Due to the cooling of the air the same pressure at the surface of the Earth over the region a will be determined by a shorter air column than over the regions which were not cooled. In cooler and denser air the pressure drops more rapidly with increasing height than in warm air. Accordingly, also the other isobaric surfaces drop over the region a. The bending of these surfaces downward means that a horizontal pressure gradient appeared, resulting in the flow of air masses from the warm regions b to the colder region a. In Diagram 3 (Figure 24) these air currents are indicated by arrows.

As a result of the arrival of new air masses at a high level over the region a the pressure at the lower level rises and, as is shown in Diagram 3 (Figure 24), the isobaric surfaces bend upward.

This indicates the appearance of a horizontal pressure gradient, directed from the increased pressure region a sideways, i.e., to the regions b. Thus the descent process of the isobaric surfaces, associated with the inflow of air at high layers and its outflow in the lower layers, arises together with the appearance of horizontal temperature differences between air masses lying side by side.

This elementary diagram gives us an idea of how air currents arise. In places where conditions for the setting up of the horizontal pressure gradient are created, air currents tending to nullify the nonuniformity of the temperature and pressure fields appear independently of the scale of this process.

Observing the weather on the sea coast, one sees that during the summer in clear and calm weather the wind directions during the day and night are different. During the day, usually pleasant fresh wind blows from the sea to the coast; at night, the wind is directed from the coast to the sea. This feature of the wind on the sea coast is due to the difference in the heating and cooling rates of the air over land and over water. During the day, the land surface, and consequently also its adjacent air layer, is heated faster than the sea and the air over it. A temperature difference is therefore created between the air masses over the land and over the sea.

Being warmer and lighter, the air over the land moves upward, and is replaced by cooler air from the sea. During the night, the air over the land cools down more rapidly than over the warm water. Air from the coast therefore moves to the sea taking the place of the warmer air ascending there.

At a certain height, air currents opposite to those which are observed at the surface of the Earth, are created. Such winds, observed on the coasts of seas and lakes, are called cold breezes.

Breezes are local winds that do not have much force. Figure 25 shows schematically the motion of air masses during the day and at night. As can be seen from the figure, the lower air current is directed from the sea during the day. This current is only several hundreds of meters high. P.A. Vorontsov and others found that at a height of 1.5–2.0 km over the Black Sea coast of Caucasia, a current arises above the breeze current and in the opposite direction. At night the directions of motion below and above are reversed. Above the breeze-circulation level the direction and velocity of the wind are determined by large-scale processes.

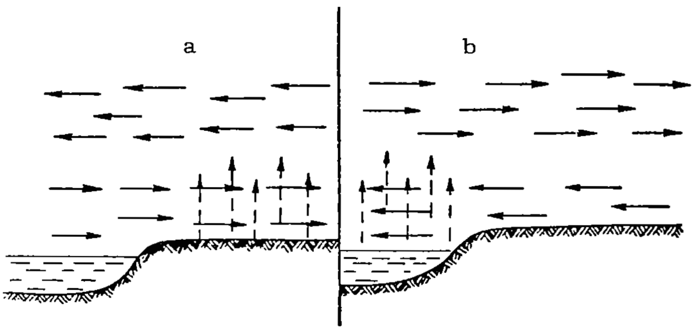

FIGURE 25. Breezes

a – during the day; b – at night.

In the case of intensive atmospheric processes, associated with strong winds, the breeze circulation, being weaker, is absorbed by the large-scale circulation.

Mountain-valley winds arise in a way similar to that of the sea breezes. During the day they are directed from the valleys upward along the mountain slopes, and at night – from the mountains to the valleys. This is caused by the unequal heating and cooling of the mountains and of the valleys. During the day the heated air over the mountain slopes has a higher temperature than the air at the same height over the valley. Being lighter, the air over the slopes moves upward, and air from the valleys moves up to the mountain slopes. At night, the mountain slopes cool down more than the air over the valleys. The cooler air from the mountains therefore flows to the valleys.

At middle latitudes, mountain-valley winds are observed in summer in calm weather up to a height of 1–3 km. Their velocities usually do not exceed 3–5 m/sec. But in rare cases, in a certain topography and state of the atmosphere, the velocity of mountain-valley winds may reach 6 m/sec and more. Mountain-valley winds are not observed in bad weather, since they are upset by the larger-scale atmospheric circulation.

A.Kh. Khrgian noticed that in the Tsei gorge (Ciscaucasia) the valley wind appears during the summer at about 0700 hrs. It attains maximum velocity at about 1500 hrs; at a height of 2 m it is equal to 2 m/sec, and at a height of 50 m – 4.2 m/sec. The valley wind in the Tsei gorge extends on the average up to a height of 1.1 km. Above this level the wind reverses its direction. At a height of 2 km the wind velocity is 1.5 m/sec.

Heating of the air over continents and oceans

Land and sea are heated unequally. The resulting differences in the air temperature and pressure distributions give rise to air currents. In contrast to the breeze winds that have a daily variation and spreading out to a short distance from the coastal region, winds arising between continents and oceans have a seasonal character and extend over large territories. They arise as follows. During the winter the air masses over the cold continents cool down and the oceans, having accumulated considerable heat reserves during the summer, lose it rapidly in heating the air. During the summer, the continents are heated more rapidly and strongly than the oceans.

Thus large and small temperature and pressure differences arise over huge areas of the continents and of the oceans, resulting in seasonal changes of the wind direction between the ocean and the land. When the land is cooled down during the winter, the wind in the lower layers converges to the warm ocean; during the summer the opposite occurs, and air flows from the cold ocean to the warm land. These seasonal winds are called *trade winds*. To compensate for the air loss in the lower layers of the troposphere, winds of opposite direction should arise in the middle and upper troposphere; this does not occur, however, since they are offset by another more powerful circulation. Trade-wind circulation extends over all regions where temperature differences arise between continents and oceans.

Many scientists have studied the trade-wind circulation. Some (S.P. Khromov, G. Flon) investigated trade winds from the point of view of the cyclonic activity and gave an explanation of this phenomenon. Others (V.V. Shuleikin, A.A. Dmitriev, A.M. Mkhitaryan and others) investigated the trade-wind circulation from a hydrodynamic viewpoint and found a number of regularities in its appearance and development.

It is very difficult to study the trade-wind circulation, since it arises on the background of a larger-scale circulation taking place between the Equator and the Poles.

The scale of the atmospheric circulation between the Equator and the Poles is even larger. As we saw above, at the Equator and at low latitudes, the inflow of solar energy is considerably larger than at middle and high latitudes. Accordingly, the heating of the masses is different at different latitudes. Cold air currents flow from high latitudes of the northern and southern hemispheres to low latitudes, and warm air currents from the equatorial and tropical zones – to high latitudes. A similar air exchange between various latitudes takes place throughout the year in both hemispheres, getting somewhat weaker during the summer and stronger during the winter. This is the basic air circulation over the terrestrial globe. The multitude of basic air currents on the Earth constitutes the *general circulation of the atmosphere*.

Air pressure and wind. Forces acting in the atmosphere upon the appearance of a wind

If the character of the air currents was determined only by the thermal nonuniformity of the surface of the Earth and of the air masses, then the regime of the wind in any part of the terrestrial globe would be simple, i.e., the wind would be determined only by the horizontal pressure gradient and the air would move along this gradient from high to low pressure. The wind velocity would then be inversely proportional to the distance between lines of equal pressure, i.e., between the isobars. The shorter the distance between the isobars, the larger the pressure gradient, and correspondingly the higher the wind velocity (Figure 26).

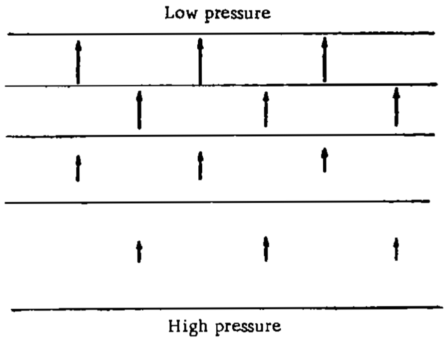

FIGURE 26. Wind direction and speed due to the force of the pressure gradient (the length of the arrow indicates the wind velocity which is proportional to the magnitude of the pressure gradient)

However, in reality the primary thermal source of air currents is coupled with the action of a whole series of other factors, thus considerably complicating the atmospheric circulation. Therefore, both the trade-wind and the interlatitudinal circulation, by means of which heat is continuously exchanged between the continents and the oceans and between low and high latitudes, respectively, are actually very complicated.

Primary among the forces causing a variation in the direction and speed of air currents is the deflecting force arising from the Earth's rotation, or, as it is usually called, the Coriolis force.

According to the law of inertia, a body maintains its state of motion if all the external forces acting on it are mutually balanced. A body moving on the rotating Earth is deflected sideways under the action of the Coriolis force, perpendicular to the direction of its relative motion.

Let us assume that at some latitude a body begins to move northward along the meridian. Maintaining its direction, the body rotates at the same time with the Earth, and after some time will therefore deviate from the direction of the meridian, since the meridian changes its direction in space and turns to the left of its initial position. Since we determine the direction of motion relative to the terrestrial surface, it seems to us that the body has deviated to the right, i.e., to the right of the initial direction of motion, not that the meridian deviated to the left. Such a deviation occurs, not only when a body moves along a meridian, but in any direction.

The Coriolis force affects all moving bodies on Earth. Its action explains the erosion of river shores. When the flow direction in a river is from north to south the right-hand shore is eroded. For this reason, the right-hand shore, as a rule, is high in contrast to the gently sloping left-hand shore. When the flow direction is from south to north the left-hand shore of the river is eroded.

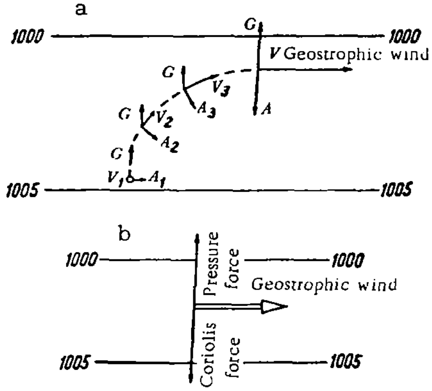

FIGURE 27. Appearance of geostrophic wind under the action of the Coriolis force

The Coriolis force also exerts an influence on the direction of sea currents, deflecting them to the right in the northern hemisphere and to the left in the southern hemisphere.

Under the action of the Coriolis force the wind does not blow along the pressure gradient but deviates from it. Air currents, beginning to move along the pressure gradient with an increasing velocity, deviate to the right of the pressure gradient in the northern hemisphere, and to the left in the southern hemisphere.

It is possible to get an idea of how the pressure force and the Coriolis force affect the direction of the real wind (there being no other forces acting) from Figure 27a. Suppose that under the action of the pressure gradient an air particle (denoted by a small circle) begins to move in the direction of the gradient. When the particle moves with the initial velocity , an acceleration A_1 arises, caused by the Coriolis force, which is directed perpendicular to and to the right of the velocity V_1. This acceleration gives the particle a velocity V_2. But then the Coriolis force varies to A_2. Under this acceleration the velocity of the air particle varies again, becoming V_3. The Coriolis force again varies, and so on. As a result, the pressure force and the Coriolis force are balanced and the air particle moves along the isobars. Such a wind is called a Gradient Wind. When moving along isobars the low pressure remains to the left of the direction of motion, and the high pressure – to the right of it. Figure 27b shows the case when there is equilibrium between the pressure force and the Coriolis force Observations show that at the height of about 1 km and higher the air moves approximately along the isobars, with small deviations caused by various reasons. With increasing latitude the Coriolis force increases, reaching a maximum at the North and South Poles. In the equatorial zone it approaches zero.

In addition to the Coriolis force, friction acts in the surface air layer directed always opposite to the direction of motion and proportional to the velocity. It reduces the velocity of the air currents, deflects them to the left of the isobars, and forces the air to flow at some angle to the isobars from the high to the low pressure.

The velocity of the moving air decreases owing to its contact with the Earth's surface. The direction also varies. Through the turbulent mixing of the air, the influence of the friction is extended to higher lying layers, up to approximately 1 km above the surface of the Earth.

The influence of friction on the direction and velocity of the air motion is shown in Figure 28 a. The diagram shows the pressure field and the motion of the air under the force of the pressure gradient, the Coriolis force, and friction. As can be seen, under the action of the Coriolis force, the flow of air takes place not along the pressure gradient G, but at right angles to it, i.e., along the isobars. The direction of the real wind is shown by the arrow W. The arrow, representing the frictional force F, is directed not exactly opposite to the wind direction, but somewhat sideways. The Coriolis force, directed at right angles to the real wind, is represented by the arrow C. As can be seen, the angle between the real wind W and the frictional force F is greater than 90°, and the angle between the real wind W and the pressure gradient force G is less than 90°. Since the gradient is perpendicular to the isobars, the real wind is deviated to the left of the isobars.

The angle between the isobar and the direction of the real wind depends on the roughness of the terrestrial surface.

Figure 28b shows the wind direction with respect to rectilinear isobars under the action of the Coriolis force and the frictional force. The wind is directed from high to low pressure. In addition, the length of the arrow indicates the wind velocity; it is proportional to the pressure gradient. The shorter the distance between the isobars, the stronger the wind. The deviation takes place to the left of the isobars usually at an angle of 20°–30°. Over land the friction is higher than over the sea. The effect of friction is highest at the surface of the Earth. With increasing height it decreases, and at the height of about 1 km the action of the frictional force almost stops.

If the isobars are curved, i.e., have, for example, the form of an ellipse or of a circle, then a centrifugal force which directs the air currents along the isobars (in the absence of friction) affects the air motion. Due to the action of the frictional force the wind blows at an angle to the isobars in the low-pressure direction. At the surface of the Earth, even over small sections, the curvilinear form of the isobars prevails.

The air pressure is determined by its mass in an atmospheric column of unit cross section. When the motion of the air is nonuniform, owing, for example, to the variation of its thermal properties, the air mass in the column decreases or increases. and the atmospheric pressure drops or rises accordingly. As a result, atmospheric vortexes – cyclones and anticyclones, often arise. In cyclones the air pressure rises from the center to the periphery, and the anticlockwise winds are directed from the periphery to the center. In anticyclones, the air pressure increases from the periphery to the center, and the clockwise winds are directed from the center to the periphery. In the southern hemisphere, the winds blow clockwise in a cyclone and anticlockwise in an anticyclone.

In addition to cyclones and anticyclones, there exist ridges, troughs, and saddle points. A ridge is a high pressure region stretching from the central part of an anticyclone with an anticyclonic system of circulation, but with open isobars.

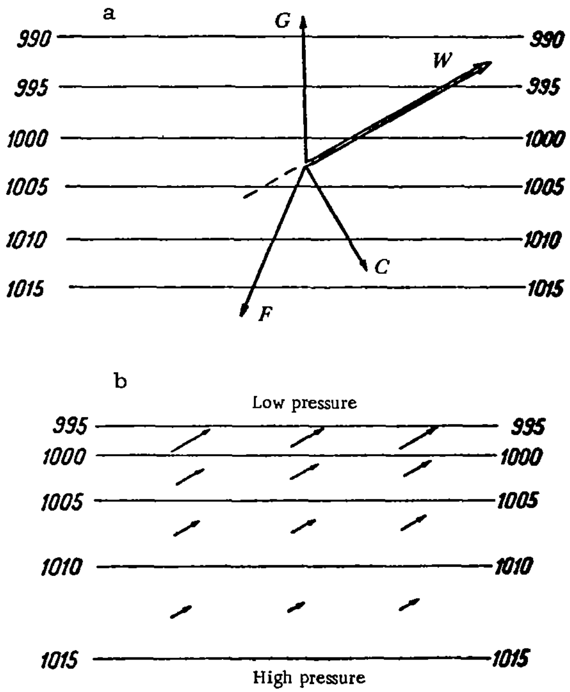

FIGURE 28. Deviation of the wind direction from the isobars under the action of the Coriolis force, the pressure-gradient force and the frictional force (a) uniform rectilinear motion in the presence of a frictional force (b)

A trough is a low-pressure region stretching from the central part of a cyclone with a cyclonic system of circulation, but with open isobars.

A saddle point is a form of a barometric topography and is situated between two cyclones and two anticyclones.

Figure 29 shows a surface pressure field with the wind system. In addition to two cyclones and two anticyclones, troughs, ridges, and saddle points are also shown. The wind direction is indicated by the arrows, the speed – by the tails. The larger the distance between the isobars, the lower the wind velocity and the smaller the tail. Such representation of wind and isobars is customary on weather maps (see below).

In spite of the development both of cyclones and anticyclones at middle latitudes of the northern and southern hemispheres, relatively low pressure prevails at the surface of the Earth. Anticyclones prevail in the subtropics. In the extreme north and south, i.e., in the Arctic and in the Antarctic, high air pressure prevails, and over the Equator – low pressure.

The atmospheric circulation over the terrestrial globe is highly diversified and complex. The regime of the air currents is different during the winter and summer at the Earth's surface and at greater heights over continents and over oceans, not to mention its large daily variations at middle

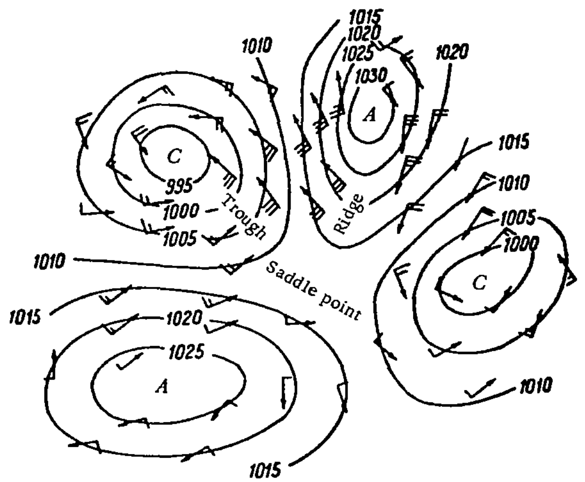

FIGURE 29. Pressure systems at the Earth's surface

C – cyclone; A – anticyclone.

and high latitudes. Usually the mean monthly maps of the pressure and of the air currents reflect only the average air-mass transfer during the month and hide many interesting features of the atmospheric processes which are observed on the daily weather maps.

Pressure and wind fields at the Earth's surface

To be able to represent the pressure distribution over the terrestrial globe or over any part of it correctly, the observational data must be comparable. Observations of atmospheric pressure are made at points which are situated at various levels; since the pressure drops with increasing height, for comparison purposes it must be reduced to one level, usually sea level. This calculation is not difficult, with the exception of special cases, since the rates of decrease of the pressure with height are known. Pressure reduction to sea level is performed not only when drawing mean maps of the pressure, but also when preparing the daily weather maps (see below).

The monthly mean maps of the pressure and air current distribution over the terrestrial globe during January and July have important differences. In the northern hemisphere during January (Figure 30) one clearly sees low-pressure regions over the warm water of the northern parts of the Atlantic and Pacific oceans, and over the cold continents (North America, Europe, and particularly Asia) – high-pressure regions. The isobars look as though they surround the continents. The so-called subtropical high-pressure belt actually has an irregular form. Whereas over the oceans

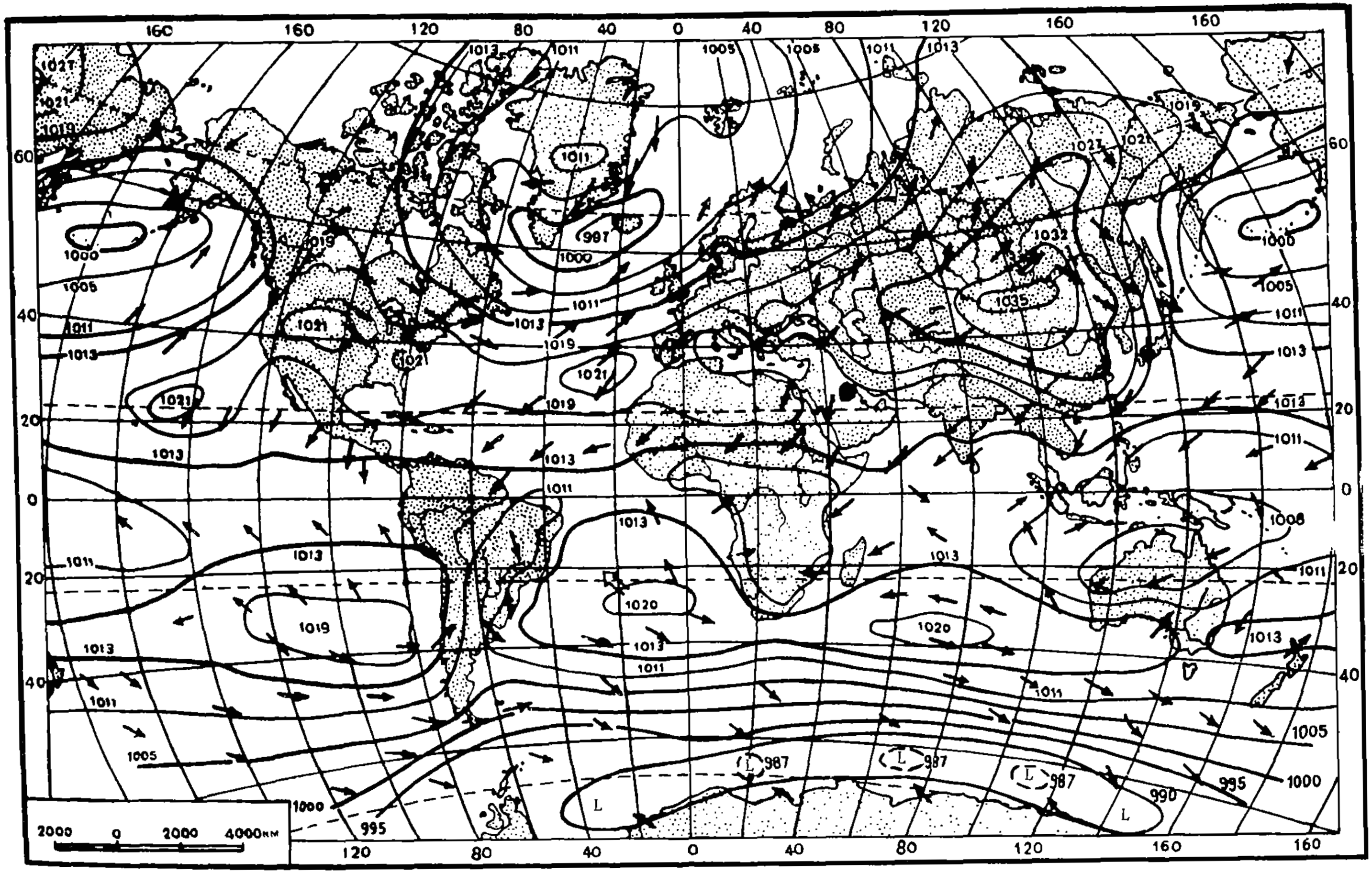

FIGURE 30. Pressure field and prevailing air currents at the Earth's surface, January

the high-pressure belt (1021 mb) is situated between latitudes 20° and 40°, over the land the center of the high-pressure region is situated over Central Asia between 40° and 60° NL, with the pressure at the center reaching 1035 mb. This distribution of the mean pressure during January points to the considerable role played, along with the cyclonic and anticyclonic activity, by the cooling of the continents, and consequently of the air masses flowing over them during the winter.

The monthly mean pressure and air-current maps, concealing many curious details, still reflect the most characteristic features of the pressure field of some regions. In particular, the low-pressure regions on the north of the Atlantic and Pacific oceans, with large pressure gradients during January reflect the often-observed cyclonic activity associated with gales, which is typical for the winter in the regions of Newfoundland, Iceland, the Norwegian Sea and the Kola Peninsula. The cyclones in the other regions of cyclonic activity – in the Far East, the Japanese islands, the north of the Pacific Ocean and Alaska are deeper and more powerful.

In contrast to these gale regions in the high-pressure regions, anticyclones with their typical low-cloudiness and calm weather extend over huge ocean spaces to the south of 40° NL. Finally, in the equatorial calm zone the air pressure and the weather conditions vary little not only daily, but also seasonally

In the southern hemisphere, where the height of the summer is January, the pressure is reduced over the warmed South America, southern Africa, and southern Australia and increased over the relatively cold oceans. Consequently the subtropical high-pressure belt with its characteristic calm weather is discontinuous in several places.

To the south of 40° SL a low-pressure belt with characteristic cyclonic activity, strong winds, and gales lies over the oceans. There is a relatively high-pressure region with very low air temperatures over the Antarctic.

The distribution of the mean pressure, and consequently of the prevailing air currents, during July, differs considerably from those of January. Although during July the isobars generally surround the continents, reduced pressure is observed in the northern hemisphere over the heated continents (Figure 31). The huge low-pressure region (1000 mb) over Eurasia, with its center over southern Asia, is particularly prominent. Consequently, the subtropical high-pressure belt is broken over the continents. It has the form of isolated high-pressure regions over the oceans with a pressure of 1027 and 1024 mb. This difference between the mean-pressure distribution in the northern hemisphere during the summer and winter is due to the large dimensions of the continents. The continents, being heated faster than the oceans during the summer, give rise to the observed pattern of the pressure distribution and have a large effect on the atmospheric processes. In the southern hemisphere where the continents occupy a relatively small area, the difference between the pressure distributions of the summer and winter is small.

During the summer in the northern hemisphere, the temperature and pressure gradients decrease and the winds become weaker compared with the winter. The cyclonic activity in the north of the Atlantic and Pacific oceans declines and gale winds become rare. In the subtropical anticyclone regions calm weather sets in. During the period of sailing boats, the calm weather at these latitudes of the Atlantic Ocean caused the death of many sailors. Sailors going from Europe to the New World were held up for a long time because there was no wind, and livestock died due to lack of food. These calm-weather latitudes thus acquired the name horse latitudes.

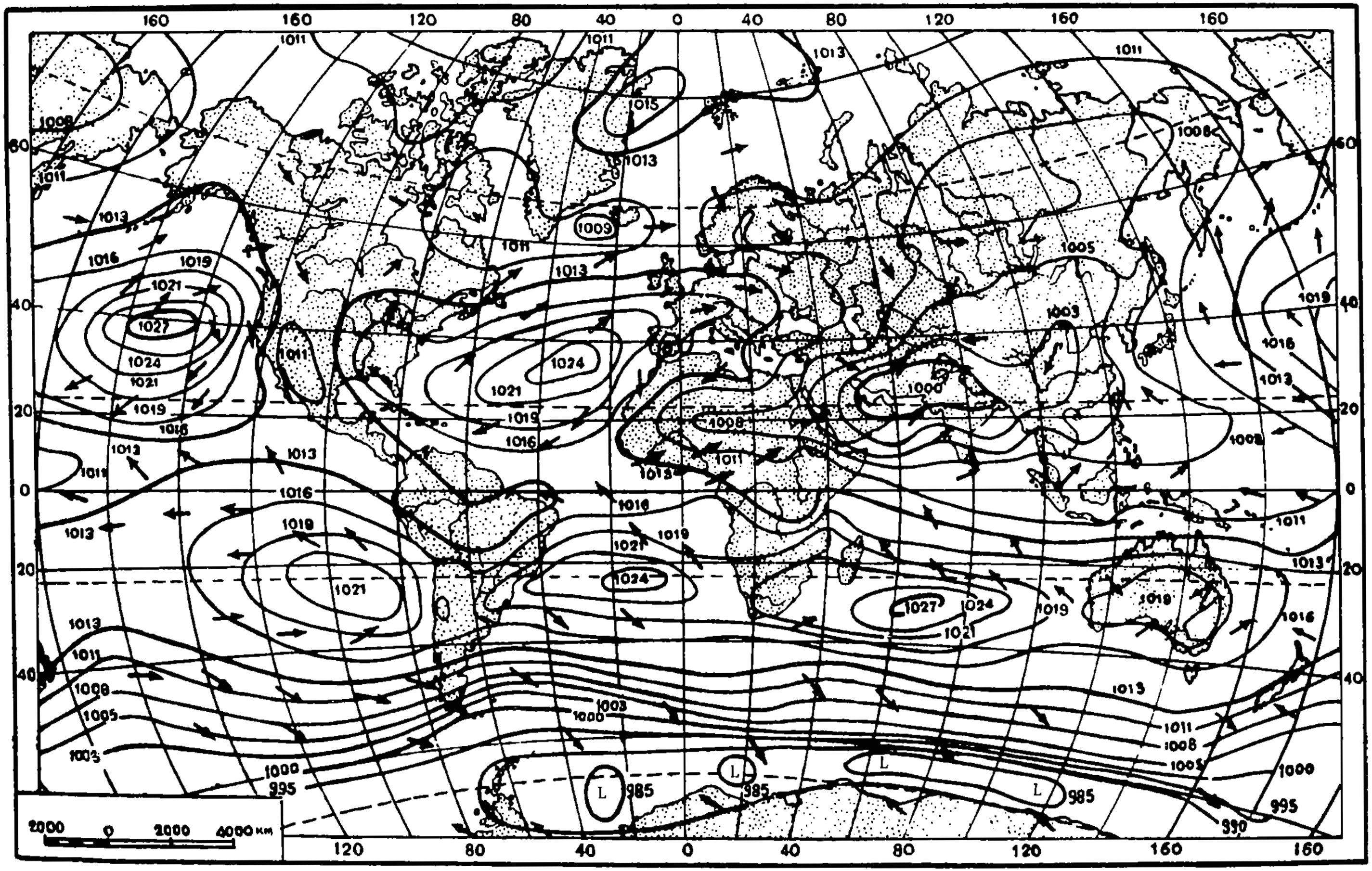

FIGURE 31. Pressure field and prevailing air currents at the Earth's surface, July

The complex atmospheric circulation can be imagined as caused by a gigantic engine with refrigerators situated at the Poles and at high latitudes, and with a powerful heater in the equatorial zone and in the tropics.

A second, less powerful engine may be imagined to operate, with the role of the heater played by the continent, and the role of the refrigerator by the ocean, or vice versa, depending on the season of the year. The work of the second engine is manifested clearly mainly at low latitudes, and this only at the surface of the Earth were the seasonal exchange of the trade wind between land and sea takes place. However, the trade-wind activity is not confined only to the tropics and the subtropics. Evidence of this may be found in the distribution maps of the mean surface pressure and of the air currents (Figures 30 and 31). On the January and July maps it is not difficult to observe a difference in the structures of the pressure and wind fields. This is clear in the northern hemisphere, where the largest continents and oceans are situated. During the winter the horizontal pressure gradients and, consequently, the prevailing winds, are directed on the average from the continents to the oceans, and during the summer, from the oceans to the continents. It is obvious that in the absence of the continents, the isobars at the surface of the Earth, and all the more so at greater heights, would surround the Earth in regular circles along the parallels. A picture close to this is observed at middle latitudes of the southern hemisphere.

We have already said that the trade-wind circulation, arising as a result of the heat exchange between the continents and the oceans, is to a considerable extent complicated by the western transfer in the troposphere. It is therefore observed only at the surface of the Earth and takes the form of a seasonal wind shift.

Depending on the season of the year, on the heating conditions, etc., the air mass over the continents and over the oceans increases or decreases. In the warm half-year the air over the continents is heated more than over the oceans. Therefore, part of the air moves from the continents to the oceans, and consequently the air mass over a unit area of the continents decreases as compared with the oceans. During the winter, the air over the continents is cooled more than over the oceans, and as a result the air mass over a unit area of the continents is larger than over the oceans. Since the air pressure depends on its mass, the pressure over the continents at the Earth's surface is higher during winter than summer.

The difference between the air pressures during January and July is considerable, a fact which can be seen from the map of the seasonal redistribution of the air masses, plotted by T.V. Bonchkovskaya and N.L. Byzove under the direction of V.V. Shuleikin (Figure 32). As shown by calculations (see the map), at a latitude of 60° over the North Atlantic the weight of an air column of cross section 1 m^2 increases from the winter to the summer by 140 kg, or by 140,000 tons per 1 km^2. Conversely, over the south of western Siberia the weight of an air column decreases by 180 kg per 1 km^2, or by 180,000 tons per 1 km^2. Considerable seasonal redistribution of the air masses takes place also at other parts of the terrestrial globe, although the largest variations are those observed in the northern hemisphere due to the large areas of the continents. Over Europe and Asia, taken together, the variation of the air weight from summer to winter reaches 5000 billions of tons. However, this impressive interseasonal difference in the weight amounts to slightly more than one-thousandth of the weight of the whole atmosphere.

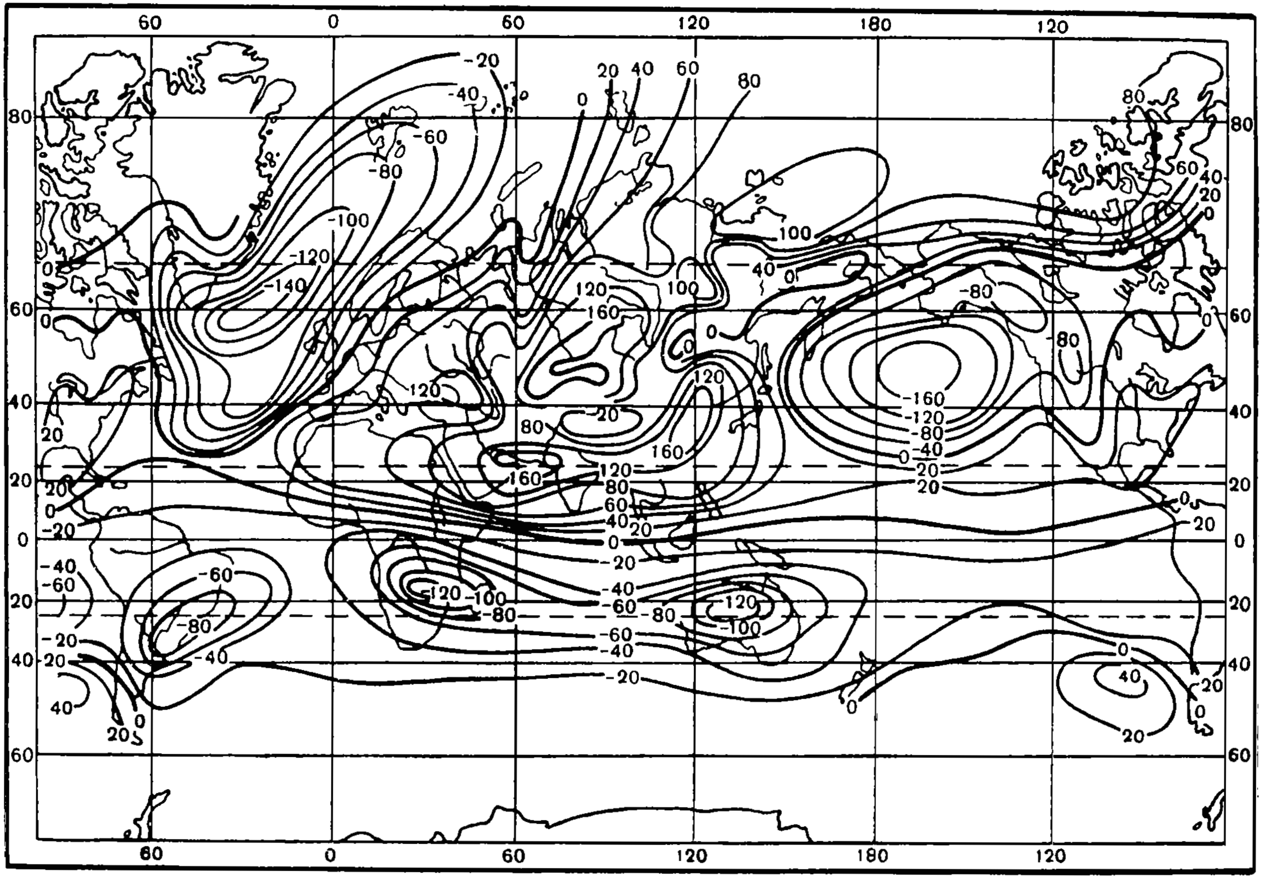

FIGURE 32. Interseasonal values for the air-mass redistributions over the terrestrial globe, January-July

From January to July and from July to January the pressure field varies gradually, preserving during the winter half-year the features of the January, and in the summer half-year the features of the July pressure field.

The pressure field is also modified with height. Whereas at the Earth's surface the continents and oceans with their seasonal warming and cooling peculiarities exert a considerable influence on the structure of the pressure field, this influence decreases with increasing height. At a height of 3 km many details of the pressure distribution and their corresponding system of winds, which is characteristic of the Earth's surface, have already disappeared. At higher levels the pressure field has features caused mainly by the difference in the inflow of solar energy at low and high latitudes of both hemispheres. Therefore, above 3 km the horizontal gradients of the temperature and, consequently, of the pressure point from low latitudes to the Poles.

As can be seen from the monthly mean temperature and pressure maps, drawn on the basis of observational data, this assumption is valid throughout the year. However, the influence of the continents and of the oceans is felt even at heights of 10–15 km.

Air currents at higher levels

The structure of the pressure field at a level of about 9 km is shown in Figures 33 and 34. These maps show the topography of the pressure field, or as it is usually called, the barometric topography (see below). These maps are therefore called barometric-topography maps. In the figures we give the topography maps of the 300-mb isobaric surface (according to the author). In these maps the isolines represent the heights of the 300-mb surface above sea level in a similar way to that in which the geographic topography is represented on a physicogeographic map. However, in contrast to the latter, the heights on the barometric-topography map are given in so-called geopotential meters (see below). The isolines are drawn through each 80 m.

The topography map of any surface of equal pressure represents the pressure field at the corresponding level. In particular, on the topography map of the 300-mb surface (Figure 33) one sees the mean distribution of the pressure and of the prevailing air currents at the level of about 9 km. It can be seen from this map that the structure of the pressure field at a height of 9 km during the winter is simpler than at the Earth's surface. From latitudes of 20°–30° both in the northern and in the southern hemispheres, the horizontal pressure gradient is directed to the Poles. The maximum wind velocities are directed from west to east along the isolines and exceed 80-100 km/hour over those parts of the terrestrial globe where the maximum density of isolines (isohypses) is observed. A sharp difference exists between the structure of the isolines in the northern and in the southern hemispheres.

In the southern hemisphere there are low-pressure troughs oriented from north to south, and two high-pressure ridges directed from south to north. The troughs are over the continents, and the ridges – over the oceans. The troughs at heights are due to the cooling of the air masses flowing over the continents and the ridges – to the heating of the air masses

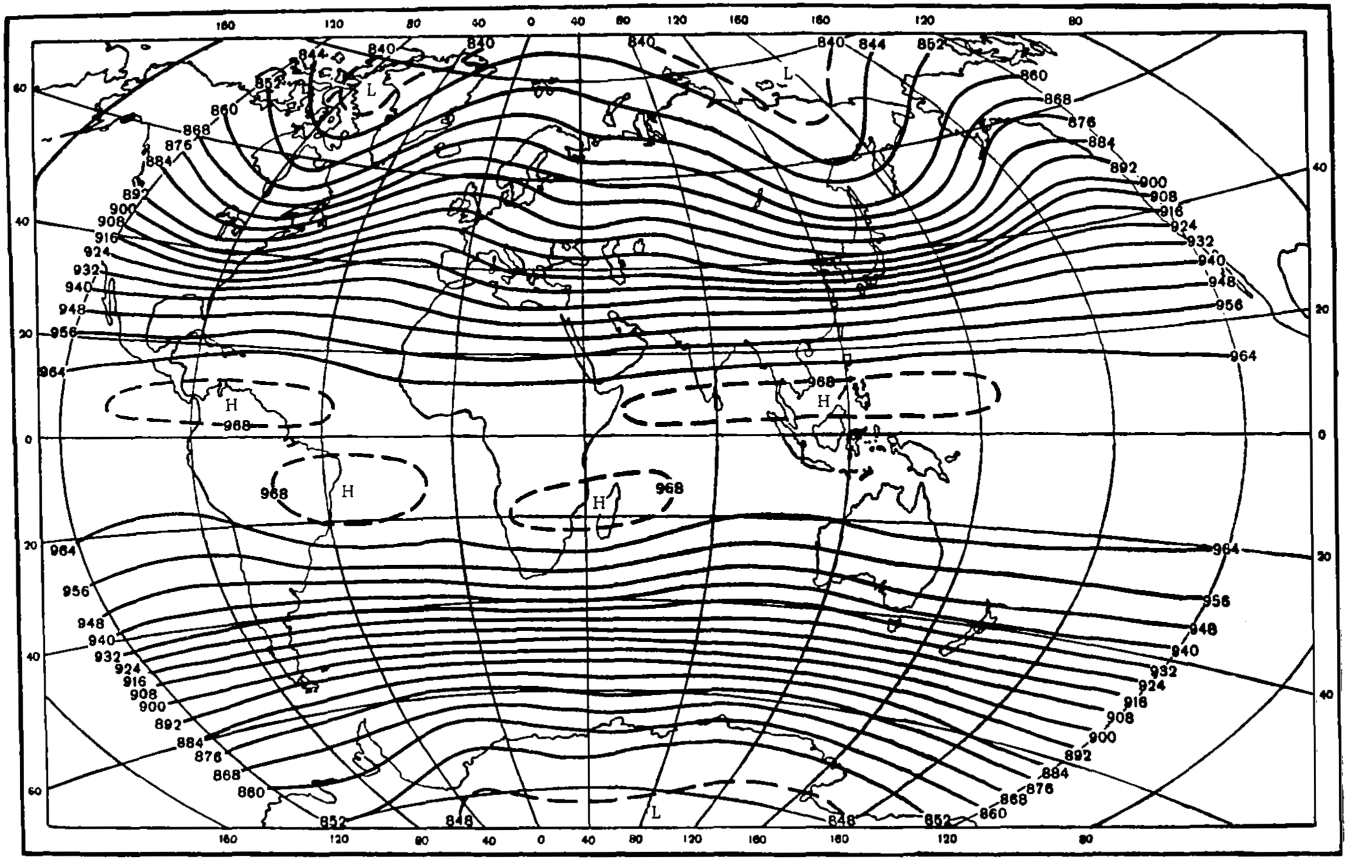

FIGURE 33. Barometric-topography map of the 300-mb surface, equivalent to the pressure field at the level of about 9 km, January

over the warm water of the oceans. The structure of the pressure field at heights shows how strongly various thermal properties of the continents and of the oceans are reflected in the system of air currents, even at a height of 9 km.

In the southern hemisphere, mid-summer is in January. There, at middle latitudes, the isolines are almost parallel, since they pass over the uniform surface of the ocean.

The pressure field at heights of from 3–5 to 15–18 km has a similar structure with the only difference that the wind velocity increases with height in the troposphere up to the stropopause. Above the tropopause, i.e., in the stratosphere, it decreases; there, the horizontal gradients of the temperature and the pressure are directed, contrary to the tropospheric case, from the Poles to the Equator. An exception are the latitudes 55°–75° in the northern hemisphere, where in the winter the temperature in the stratosphere drops with increasing height at high latitudes.

During July in the northern hemisphere the structure of the isohypses is considerably different. During the summer, as a result of the temperature rise in the stratosphere at high latitudes, the pressure gradients, and consequently also the wind velocities, decrease. In addition, the troughs and ridges almost disappear, since the temperature over the continents and over the oceans is almost equalized (Figure 34).

In the southern hemisphere during July, i.e., mid-winter, the pressure gradients are somewhat larger than during January, but the general configuration of the isohypses does not vary appreciably.

However, in spite of the seasonal variations, the zones of largest temperature and pressure gradients span the northern and southern hemispheres throughout the year. These zones are called *planetary high-level frontal zones**.

The highest wind velocities occur in the planetary frontal zones. High-level mean pressure maps smooth out many interesting and important features of the atmospheric circulation. For example, the heat and moisture transfer from low to high latitudes was almost absent from the monthly mean pressure maps. Yet this daily interlatitudinal exchange of air masses provides heat and moisture to the middle and high latitudes of both hemispheres.

Heat exchange between low and high latitudes

The basic mechanism of interlatitudinal air transfer is the cyclonic and anticyclonic activity, which transfers warm and humid air masses from the south to the north and cold and dry masses from the north to the south in the northern hemisphere.

Owing to the position of the continents and oceans, the most powerful interlatitudinal exchange of air masses takes place in the northern hemisphere. As a result, the temperature in the Arctic often rises to 5°–10°C during the summer and in individual cases to 15°–20°C above zero. During the winter thaws are often observed in the Atlantic sector of the Arctic, in the region of Spitsbergen and of Franz Josef Land.

* For more details see the book: Pogosyan, Kh.P. Planetarnye frontal'nye zony v severnom i yuzhnom polushariyakh (Planetary Frontal Zones in the Northern and Southern Hemispheres). —Leningrad, Gidrometeoizdat. 1955.

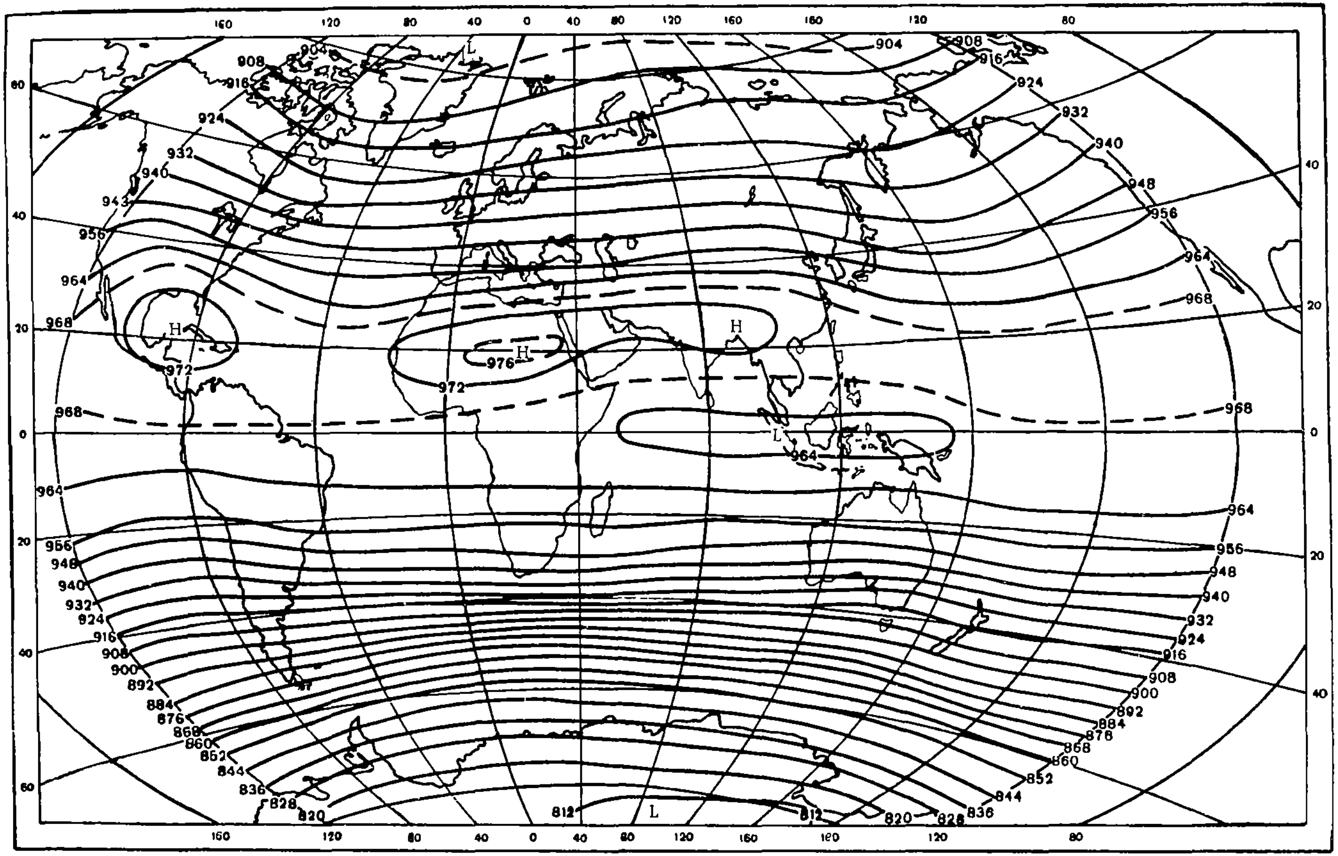

FIGURE 34. Barometric-topography map of the 300-mb surface, equivalent to the pressure field at the level of about 9 km, July

In the southern hemisphere the interlatitudinal air exchange is also intensive. However, in the Antarctic it does not reach the same dimensions that are characteristic of the Arctic. This is because at middle latitudes of the southern hemisphere there are no alternating continents and oceans as in the northern hemisphere. In addition, Antarctica is almost symmetrical with respect to the South Pole. Cyclones and anticyclones are therefore not localized there as in the northern hemisphere, since conditions are not created for a prolonged transfer of warm air. Air masses moving from the north are cooled over ice-covered Antarctica, thus reducing the scale of the interlatitudinal air exchange at high latitudes of the southern hemisphere.

Another important transporter of heat from low to high latitudes is the ocean (see the chapter "The Air Temperature at the Surface of the Earth").

Warm currents temper the climate of cold countries appreciably during winter, and cold currents moderate the tropical heat at low latitudes. Heated during the warm half-year, the ocean loses this heat during the winter on heating the air masses flowing over it.

Inhabitants of the western and central regions of the European territory of the USSR sometimes observe a temperature rise during the winter and even thaws. The temperature rise is usually connected with the intensification of the warm air transport either from the far south beyond the borders of the Soviet Union, or from the Atlantic Ocean – the powerful heat center of Europe and of part of Asia.

V.V. Shuleikin calculated the amount of heat being transferred across the Atlantic coast of Europe, and showed for the first time the immense role of the oceanic heat transfer in the thermal balance of the northern seas of the USSR.

Later, on the basis of data of aerological observations, the author established that heated and moistened masses of air with a vertical thickness of over 5–7 km, moving over land from the coast of the Atlantic Ocean, not only warm the whole of Europe but exert a definite influence on a considerable part of North Asia, up to Yakutia. The heat is transported from the west not only over land, but also over the northern seas. For just this reason strong frosts in the north of the European territory of the USSR are frequently replaced by warm weather, the heat arising not from the south or from the west, as usual, but from the direction of the Arctic seas in the north. The air heated by the ocean reaches far over to the east, particularly over the north of the Asian Continent.

The influence of the Atlantic Ocean on the climate of Asia is considerably larger than that of the Pacific Ocean. This is due to the prevailance of the western transport of air masses in the troposphere. Considering that the Pacific Ocean is such a huge heat reservoir, the Asian Continent receives a negligible amount of heat. It is shown by observational data that the mean temperature for January in Central Europe at 50° latitude is approximately 1°C below zero, while in eastern Asia at the same latitude and at the same distance from the shores of the Pacific Ocean, the January temperature is about 20°C below zero.

Because of the prevailing of the western transport of air masses in the troposphere, the warm waters of the North Atlantic, softening the winter climate of Europe, exert only a small influence on the climate of all of North America, although this continent is not farther from the warm currents of the Atlantic Ocean than Europe.

Thus, owing to the continuous air motion, heat and air are exchanged between the Poles and low latitudes, and between continents and oceans. The moisture evaporating from the oceans and from the continents is transported by the air currents both in the horizontal and vertical directions.

The nonuniform distribution of the precipitation over the terrestrial globe is determined by the atmospheric circulation. It gives rise to droughts and dry winds, to gales and snowstorms as well as to other atmospheric phenomena. The expected variations in the nature of the air currents, at present often incorrectly determined, are important for weather forecasting.

The destructive force of the wind

The wind often reaches such a tremendous speed that it causes colossal damage. Such winds are often observed on the northwest and north coasts of the Pacific and Atlantic oceans. In order to reduce the destructive force of the wind, the maximum possible wind velocities at a given place are taken into account in the construction of houses and industrial plants, and in the erection of towns and settlements. The effect of the wind is taken into account in artillery operations, bomb dropping, anti-chemical warfare defences, flight routes of airplanes, etc.

The harmful action of the wind is manifested in many forms. By lifting dust from dry soil the wind creates dust storms. It also causes large losses to agriculture by blowing away the soil layer that is rich in mulch.

Winds of over 30 m/sec are often observed in a system of deep cyclones or on the periphery of powerful anticyclones. Depending on the topography, the wind may decrease or considerably increase. Such a wind is the cold northerly wind in Novorossisk, arising from the penetration of cold air from the north to the south of the European territory of the USSR and to the Ciscaucasia during winter. Air masses of a higher density than those over the Black Sea cave in from the nearest mountains of the Novorossisk Bay. Often the cold wind, descending with a velocity of up to 40 m/sec lifts as it hits the surface of the sea a huge amount of spray, which freezes at low temperatures, covering vessels and other objects with a thick layer of ice.

Water spouts or sandstorms, which are vortexes with a vertical axis, attain a destructive force. Their diameter at the surface of the sea usually reaches up to 20–100 m, and over land – up to 1–2 km (Figure 35). The wind velocities in a water spout or sandstorm reach 100 m/sec. These velocities are sufficient to destroy age-old forests, villages etc. In a system of such vortexes the air often moves counterclockwise. They arise in a thick thunderstorm cloud, from which a kind of funnel with a spout at the foot comes down. If the water spout reaches the sea, then this funnel, touching the water, expands and begins to suck water upward. Thus in water spouts forming over water, fish, lifted together with the water, may fall from the clouds.

When a vortex forms, the air on the neighboring sections descends, thus causing the vortex to close; due to the high angular velocity inside the vortex, a centrifugal force arises and the air pressure drops. That is why water or other objects are sucked in and then thrown out on the path of the water spout. *

Water spouts or sandstorms are often observed in North America where they are called tornadoes, which in Spanish means "rotating."** In the Soviet Union they do not occur often, but are usually seen once a year or so.

FIGURE 35. Water spout on the sea

The tropical cyclones, called in the Far East typhoons, have a tremendous destructive force. They are of large diameter and vertical extension, and are described in the chapter dealing with cyclones and anticylones.

Wind — the most important source of energy

Weather forecasting is not the only reason for studying the motion of the air. The wind is one of the most important sources of energy. This destructive force can be transformed into a useful one for man.

The energy resources of the wind are enormous. They are about five thousand times the annual amount of coal and gas used by man for energy.

* See the interesting book: Kolovkov, N.V. Grozy i shkvaly (Storms and Squalls). — Moskva, Gostekhizdat. 1954.

** [Sic. The origin is in fact the Spanish "tronada," which means "thunderstorm."]

Moreover, like the hydroenergy reserves, the energy of the wind is practically inexhaustible. The inexhaustibility of the energy reserves of the wind lies in the constant process of nonuniform heating of the Earth, giving rise to air currents. Although the wind energy has been used from ancient times (windmills, sailing ships, etc.), the technical means for using the wind are far from perfect. Even today wind energy is hardly used. The theory of the modern wind engine was formulated by the founder of aerodynamics and aeronautics N.E. Zhukovskii.

Research is being carried out in the Soviet Union toward producing technically-improved and economically-advantageous electric-wind engines. The first electric-wind station, which supplied 100 kw of energy, was constructed in Crimea in 1937. There are also electric-wind engines designed by V.P. Vetchinkin, A.G. Ufimtsev and others.

Electric-wind engines are used in rural areas for illuminating houses, irrigating fields, and supplying power to workshops, sawmills and so on. Electric-wind engines are particularly advantageous in regions where the wind velocities are high throughout the year. Wind energy is very important in those regions situated far from deposits of oil, coal and other types of energy.

THE VERTICAL TEMPERATURE LAPSE RATE

In the first chapter we discussed the vertical structure of the atmosphere and, to some extent, the temperature variation with height. We now consider some interesting features of the temperature regime at heights. We recall that in the troposphere the temperature decreases with height by, on the average, 0.5°–0.6°C per 100 m, or 5°–6°C per km. The temperature variation per 100 m of height is called the vertical *temperature lapse rate*.

The lapse rate is not constant, but undergoes variations for a number of reasons, thereby very often deviating from the above-indicated mean value. The lapse rate is different during the winter and summer, at night and by day, over sea and over land. This variation is particularly characteristic of the lower air layers of up to 1–2 km thickness, but it also occurs at great heights.

Even in the troposphere the temperature often does not drop but rises with increasing height. Thus an airplane may happen to reach an air layer with a higher temperature than that at the surface of the Earth. However, the temperature in the troposphere usually drops with height, since the lower air layers are heated by the surface of the Earth. The larger this heating, the higher the lapse rate in the lower layers of the troposphere. The lapse rate in the south is therefore particularly high during the summer when the heating of the terrestrial surface is most intensive. In the lower air layer during the summer, the lapse rate frequently exceeds 1°C per 100 m.

During the winter the picture is reversed. Due to the cooling of the surface of the Earth and the adjacent air layers, the temperature rises with height on land. This is because the air masses situated in higher layers do not have time to cool down to the same extent as those closer to the surface of the Earth. This condition is called *temperature inversion*.

The strongest inversions appear during the winter in Siberia, particularly in Yakutia, where at that time of the year the weather is clear and calm. Under these conditions the air is cooled by the underlying surface for a prolonged time. Therefore inversion is very often observed up to a height of 2–3 km. During the winter in Siberia, North Canada, and the shores of Antarctica, at a temperature of -50° to -60°C at the ground level the temperature at the upper boundary of the inversion reaches -30° to -35°C. Thus the temperature difference between the lower and upper limits of the inversion may amount to 20°–25°C.

The lapse rate usually varies over a 24-hour period. Due to the daily heating and to the night cooling, the lapse rate in the first 1.0–1.5 km above the surface of the Earth undergoes daily fluctuations. By day large values of the lapse rate are usually observed in this layer, increasing until afternoon; toward evening the lapse rate gradually decreases, and at night inversion often appears.

Stable and unstable atmosphere

In addition to the above-mentioned temperature variation with height, related to the heat inflow or outflow, the air temperature also depends on the atmospheric pressure. The temperature changes, for example, as a result of vertical motion of the air, when the ascent and descent take place without exchange of heat with the surrounding medium, i.e., adiabatically. During its ascent the air expends energy (from its internal energy) in order to overcome the external resistance. Therefore rising air cools down, while descending air heats up. The adiabatic temperature variations follow the dry-adiabatic or wet-adiabatic curves.

Dry-adiabatic lapse rate is the decrease in temperature of unsaturated air per 100 m of ascent.

Wet-adiabatic lapse rate is the temperature decrease of saturated air per 100 m of ascent.

Theoretical calculations, as well as observational data, show that during the ascent of dry air or of unsaturated air, the temperature decreases according to the dry-adiabatic lapse rate, i.e., by 1°C per 100 m of ascent. In descending, the temperature rises by 1°C per 100 m. This value of the dry-adiabatic lapse rate does not vary as long as the air in the ascent does not reach saturation. The level at which the air becomes saturated is called the condensation level. Upon reaching this level condensation of the water vapor begins and the latent heat of vaporization is released, thus heating the air. This additional heat in a wet-adiabatic process reduces the cooling, and therefore during the ascent of saturated air its temperature drops with increasing height by less than 1°C per 100 m. It is obvious that the higher the moisture content of the saturated air, the more latent heat is released for the same amount of cooling, and the lower the wet-adiabatic lapse rate.

At low temperatures, and consequently when the moisture content of the air is low, the wet-adiabatic lapse rate is close to the dry-adiabatic one. Thus, for example, at a temperature of -40°C in the lower half of the troposphere the wet-adiabatic lapse rate is 0.97°C per 100 m. At high temperatures, when the moisture content of the air is higher, the wet adiabatic lapse rate is considerably lower. Thus, for example, for an air temperature of 35°C it is approximately 0.3°C per 100 m.

Table 19 gives average values of the wet-adiabatic lapse rate for various temperatures and for the pressures of 1000, 750, and 500 mb, corresponding approximately to ground level, and to heights of 2.5 and 5.5 km, respectively.

TABLE 19

Cooling rates of saturated air for each 100 m of ascent

Pressure, mb	Temperature, °C						
	30	20	10	0	-10	-20	-30
1000	0.36	0.43	0.53	0.65	0.76	0.87	0.92
750	0.33	0.39	0.48	0.59	0.71	0.84	0.90
500	0.29	0.33	0.41	0.51	0.64	0.78	0.87

The wet-adiabatic lapse rate also depends on the air pressure; the lower the air pressure, the lower the wet-adiabatic lapse rate for the same initial temperature. This is due to the fact that at a low pressure the air density is also lower; consequently, the condensation heat released is used up in heating a smaller air mass.

During the summer, the lapse rate is on the average 0.6°–0.7°C per 100 m of ascent. The approximate values of the temperature at various heights can be calculated from the temperature at the ground level. If, for example, the temperature at the ground level is 28°C, then, taking an average lapse rate of 7°C per km, one gets a temperature of 0°C at a height of 4 km. The lapse rate over land during winter in the Soviet Union rarely exceeds on the average 0.4°–0.5°C per 100 m. In individual air layers the temperature sometimes almost does not vary with height, i.e., isothermal conditions prevail.

The equilibrium condition of the atmosphere – stable or unstable – can be assessed from the magnitude of the lapse rate.

In the case of stable equilibrium air masses do not tend to move vertically. In this case, if an air volume is displaced upward, it will return to its initial position.

Stable equilibrium exists when the lapse rate of unsaturated air is lower than the dry-adiabatic lapse rate, and the lapse rate of saturated air is lower than the wet-adiabatic lapse rate. If in this case a small volume of unsaturated air is raised by an outside force to some height, then as soon as the action of the external force stops, the air will return to its previous position. This is because the lifted air volume, having spent internal energy on its expansion, cools down by 1° per 100 m (according to the dry-adiabatic lapse rate). Since the lapse rate of the surrounding air was lower than the dry-adiabatic one, the air volume lifted to this height had a lower temperature than the surrounding air. Having a higher density, the volume then descends until it reaches the initial state. Let us illustrate this by an example.

Assume that the air temperature at ground level is 20°C, and that the lapse rate is 0.7°C per 100 m. For this lapse rate the air temperature at a height of 2 km will be 6°C (Figure 36 a). A volume of unsaturated or dry air raised by an external force from the surface of the Earth to this height, cooling down according to the dry-adiabatic lapse rate, i.e., by 1°C per 100 m, cools down by 2°C thus reaching 0°C. This air volume is 6°C colder and consequently also heavier than the surrounding air. It thus begins to descend, tending to reach the initial level, i.e., ground level.

A similar result is obtained in the case of ascent of saturated air, if the lapse rate of the surrounding medium is lower than the wet-adiabatic lapse rate. Therefore, in a homogeneous mass of air under a stable state of the atmosphere, no stormy formation of cumulus and cumulonimbus clouds takes place. However, in the case of frontal zones and the associated ascending air motion, layer clouds and precipitation are formed.

The most stable state of the atmosphere is observed when the lapse rate is small and particularly in inversions, since in this case warmer and lighter air lies over the lower cold, and consequently, heavier air.

In the case of unstable equilibrium, an air volume lifted from the ground level does not return to its initial position, but maintains an upward motion until its temperature becomes equal to the temperature of the surrounding air. An unstable state of the atmosphere is characterized by a high lapse rate, which is caused by the heating of the lower air layers. In this case the air masses, heated from below, move upward and the upper cold air masses move downward. This process continues until a stable equilibrium is reached. This is how thermal convection arises.

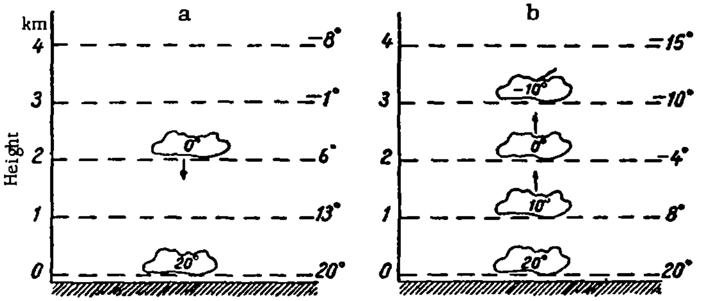

FIGURE 36. Vertical motion of an air volume

a – in the case of stable equilibrium; b – in the case of unstable equilibrium.

Suppose, for example, that unsaturated air in the lower layers has an unstable stratification, i.e., up to a height of 2 km its temperature decreases with height by 1.2°C per 100 m. Ascending in such a medium according to the dry-adiabatic law, an air volume with the temperature of 20°C at the ground level, will have a temperature of 10°C at 1 km, i.e., it will be 2°C warmer than the surrounding medium (Figure 36 b). Being also lighter, this volume moves upward. At a height of 2 km it is already 4°C warmer than the surrounding medium, since its temperature reached 0°C, whereas the temperature of the surrounding air is -4°C. Still being lighter, this volume rises and reaches a height of 3 km. Since in this case the temperature of the surrounding air in the 2–3 km layer drops by only 0.6°C per 100 m, the temperature difference between the ascending air volume and the surrounding medium disappears. Thus the free ascent of the air volume will stop.

If the lapse rate of dry or unsaturated air is lower than the dry-adiabatic one, the air is in a stable condition. If, however, the lapse rate is higher than the dry-adiabatic, the air is in an unstable condition. The state of saturated air can be determined similarly. In this case the lapse rate should be compared with the wet-adiabatic lapse rate for the given temperature and pressure.

If the lapse rate is lower than the wet-adiabatic one, the state of the air is stable; if, however, the lapse rate is higher than the moist-adiabatic rate, the state of the air is unstable. Depending on the degree of stability, good-weather cumulus clouds or thick cumulus and cumulonimbus clouds are formed. Showers, storms, squalls, and other weather phenomena result from a state of instability.

The so-called airplane bumps, in which the airplane is thrown up and down, or, as is usually said, the airplane falls into "air pockets," are caused by powerful vertical motions developing in an unstable air. At daytime, particularly during a summer afternoon, when due to the heating of the lower air layer considerable instability develops, airplane bumpiness is strongest. Night flights are usually calmer than daytime ones.

Examples of temperature variation with height

The temperature at higher elevations is very diverse. It depends on the character of the underlying surface, the time of the year, and the horizontal and vertical circulation.

Tables 20, 21, and 22 give data on the variation of the temperature with height on individual days. These data were obtained by radiosonde techniques.

Over Lvov, Arkhangelsk, and Moscow there is a uniform temperature drop with height up to the tropopause, with the lapse rate varying from 0.4°–0.9°C per 100 m (see Table 20). In the tropopause and above, the temperature does not decrease with height. In Lvov this change occurs at a height of 6.7 km, in Arkhangelsk – at a height of 8.4 km, and in Moscow – at a height of 8.5 km. Moreover, over Arkhangelsk and Moscow the temperature above 8.7 km rises somewhat, i.e., a weak temperature inversion exists; this is characteristic of the lower layers of the stratosphere.

The temperature distribution with height over Moscow, Gorki, and Minsk somewhat differs from the previous examples (see Table 21). In all three cases, temperature inversion is observed. Over Moscow it exists in the layer between the ground level and the height of 2.0 km. The lapse rate varies there from 0.0°–0.8°C per 100 m, i.e., during an ascent of 100 m the temperature rises from 0° to 0.8°C. In Gorki the same picture holds. In the layer 1.0–1.1 km the temperature rises by 1.4°C. Over Minsk in the layer 0.9–1.8 km isothermal conditions prevail. Isothermal and inversion conditions, indicating the transition from the troposphere to the stratosphere, are observed over Gorki at a height of 8.7 km, and over Minsk – at a height of 11.3 km.

Finally, Table 22 gives data on the variation of the temperature with height in the presence of surface inversions. As can be seen from this table, in Novosibirsk in the layer between the ground level and the 1.3-km level, the lapse rate reaches 1.3°C per 100 m. In otherwords, at heights of 1.3-1.7 km the air temperature is 15.8°C higher than at the ground level.

In Tambov the temperature difference between the ground level and the height of 1.4 km is 16.3°C. For a temperature of -15.5°C near the ground level, a temperature of 0.8°C is recorded at a height of 1.4 km. Above this level the usual temperature drop with increasing height takes place, and near the tropopause, at a height of 11.7 km, it reaches -59.5°C. As was mentioned above, such temperature variation with height is due to the cooling of the lower air layers; this occurs because they are in contact with the cold underlying surface.

In order to represent the temperature variation with height, special forms of aerological diagrams are usually used. On these temperature and pressure scales, the system of dry and wet adiabatic lines, and lines of equal specific humidity are reproduced; but, as before, we shall use a simpler graph (Figure 37). From Figure 37, curve 1, we see that during the morning of 4 February, 1947 in Yakutsk, the temperature rose with height up to 2.1 km. Whereas at the surface of the Earth the

TABLE 20

Temperature variation with height in the troposphere

	Height above sea level, *km*										
	Ground level 0.325	1.4	2.3	2.9	4.5	5.3	6.2	6.7	8.7	11.4	11.8
	Lvov, 0500 hrs, 11 March, 1953										
Temperature, °C	−4.5	−11.2	−17.8	−20.8	−32.0	−38.0	−45.0	−0.48	−0.48	−0.48	−0.48
Lapse rate, °C/100 m	—	−0.6	−0.7	−0.6	−0.7	−0.8	−0.8	+0.6	0.0	0.0	0.0

	Height above sea level, km									
	Ground level 0.006	1.4	2.8	5.2	7.8	8.4	8.6	8.7	10.4	11.4
	Arkhangelsk, 0500 hrs, 26 May, 1953									
Temperature, °C	1.5	−8.5	−19.2	−33.8	−42.0	−47.0	−45.0	−44.8	−42.0	−41.8
Lapse rate, °C/100 m	—	−0.7	−0.8	−0.6	−0.3	−0.8	+1.0	+0.2	+0.2	+0.0

	Height above sea level, km													
	Ground level 0.16	1.0	1.4	2.7	2.9	4.0	5.4	6.1	7.0	8.5	8.6	9.0	9.6	11.7
	Moscow, 1100 hrs, 26 May, 1953													
Temperature, °C	7.2	−0.2	−4.3	−14.8	−16.3	−21.2	−28.8	−33.0	−37.4	−43.0	−43.0	−42.2	−41.0	−41.2
Lapse rate, °C/100 m	—	−0.9	−1.0	−0.8	−0.7	−0.4	−0.5	−0.6	−0.5	−0.4	0.0	+0.2	+0.2	−0.0

TABLE 21

Temperature variation with height in the troposphere

	Height above sea level, km										
	Ground level 0.16	0.8	1.2	1.4	2.0	3.0	4.3	5.0	5.6	8.0	10.0
	Moscow, 1100 hrs, 11 November, 1952										
Temperature, °C	—4.8	1.8	1.8	3.5	3.5	—2.5	—11.2	—12.8	—17.2	—34.8	—49.0
Lapse rate, °C/100 m	—	+0.5	0.0	+0.8	0.0	—0.6	—0.7	—0.2	—0.7	—0.7	—0.7

	Height above sea level, km									
	Ground level 0.16	0.8	1.0	1.1	2.5	2.9	5.3	8.1	8.7	11.3
	Gorki, 0500 hrs, 30 December, 1952									
Temperature, °C	—9.8	—13.0	—13.2	—11.8	—11.8	—15.8	—33.5	—56.0	—57.2	—57.2
Lapse rate, °C/100 m	—	—0.5	—0.1	+1.4	0.0	—1.0	—0.7	—0.8	—0.2	0.0

	Height above sea level, km													
	Ground level 0.21	0.9	1.6	1.8	2.3	3.1	3.2	4.0	5.1	5.7	9.0	9.3	11.3	12.3	12.8
	Minsk, 1700 hrs, 3 April, 1953														
Temperature, °C	7.3	3.8	3.8	3.8	—0.8	—4.5	—5.2	—8.2	—14.5	—19.5	—48.0	—49.8	—62.0	—62.0	—57.0
Lapse rate, °C/100 m	—	—0.5	0.0	0.0	—0.9	—0.7	—0.7	—0.4	—0.6	—0.8	—0.9	—0.6	—0.6	0.0	+1.0

TABLE 22

Temperature variation with height in the case of an inversion

	Height above sea level, km										
	Ground level 0,47	0.5	1.0	1.5	2.4	3.2	3.8	4.2	5.6	6.1	6.5
	Irkutsk, 0700 hrs, 27 February, 1947										
Temperature, °C	−14.0	−12.9	1.7	−0.7	−9.3	−17.4	−23.9	−27.7	−32.5	−33.4	−40.9
Lapse rate, °C/100 m	—	+3.7	−2.9	−0.5	−1.0	−1.0	−1.1	−1.0	−0.3	−0.2	−1.9

	Height above sea level, km							
	Ground level 0,13	1.3	1.7	3.0	5.0	7.0	8.0	9.6
	Novosibirsk, 0800 hrs, 16 February, 1947							
Temperature, °C	−33.4	−17.6	−17.6	−22.1	−33.1	−42.5	−47.0	−54.3
Lapse rate, °C/100 m	—	+1.3	0.0	−0.3	−0.5	−0.5	−0.4	−0.4

	Height above sea level, km													
	Ground level 0,14	0.3	0.5	0.9	1.4	1.5	2.0	3.0	4.0	5.6	6.4	9.1	10.0	11.7
	Tambov, 0500 hrs, 26 December, 1952													
Temperature, °C	−15.5	−13.8	−9.0	−0.8	0.8	0.5	−0.8	−7.2	−12.0	−23.5	−30.0	−50.5	−58.0	−59.5
Lapse rate, °C/100 m	—	+1.1	+2.4	+2.0	+0.3	−0.3	−0.3	−0.6	−0.5	−0.7	−0.8	−0.8	−0.8	−0.1

temperature was 34.5°C below zero, at a height of 2.1 km it reached 13.5°C below zero, i.e., at this height the temperature was 21.0°C higher than at ground level. Here, as in the cases given in Table 19, the temperature rise with height was caused by intensive cooling of the air layers adjacent to ground level. Such a temperature variation with height is observed more often during the winter inside the continents in calm anticyclonic weather, when the mixing of the air is weakest.

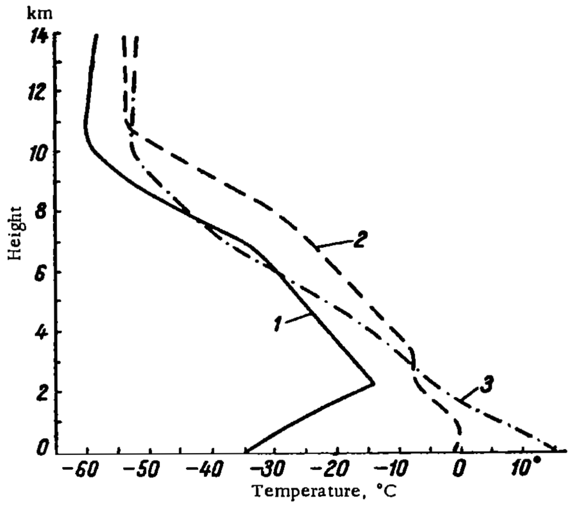

FIGURE 37. Variation of temperature with height over various places in Europe and Asia

1 - Yakutsk, 4 February, 1947; 2 - Minsk, 29 October, 1952; 3 - Berlin, 26 October, 1952.

Curve 2 represents the variation of temperature with height over Minsk during the morning of 29 October, 1952. From ground level to a height of 1 km, the temperature rose slightly due to the night cooling of the surface air layer. Above 1 km, isothermal conditions were observed in a layer about 1 km thick after the usual drop in temperature with height. This indicates the presence over Minsk of a surface of separation between air masses of different properties (see next chapter). Finally, curve 3 shows that over Berlin during the morning of 26 October, 1952 the temperature dropped continuously with increasing height. In the layer between ground level and the 10.5-km level the temperature dropped by almost 70°C, from 15° to -55°C, thus yielding a lapse rate of approximately 6.7°C per 1 km of ascent.

Judging from the character of the stratification curves, the tropopause is situated at heights of 10–11 km. There the temperature stops dropping with height. Above the tropopause lies the stratosphere, in which the temperature undergoes only small variations with increasing height.

The regional temperature distribution in the free atmosphere

The annual variation of the temperature at higher elevations with latitude over ocean and land is similar in general to the annual variation of the temperature at the surface of the Earth. The smallest temperature differences between winter and summer months are observed at low latitudes, the largest differences — at middle and high latitudes. At the same latitude, the smallest differences are observed over oceans, the largest differences — over land areas far from the oceans.

The above-given curves of the temperature variation with height (Figures 20, 21, 37) show the peculiarities of the temperature distribution with height in various regions.

Since the temperature in the stratosphere is determined mainly by the influence of the radiation heat exchange, it has a clear seasonal variation, while along the parallels during a given season, the temperature is approximately the same.

In the troposphere the temperature along the parallels is not uniform. For example, in Reykjavik (Iceland) the difference between the mean January and June temperatures at ground level is about 11°C, increasing at the 5-km level to 14°–15°C. At approximately this latitude in Yakutsk the difference between the mean temperatures of these months at ground level is the largest, reaching 33°C, whereas at a height of 5 km it is 16°–17°C. This is because Reykjavik is situated to the north of the Atlantic and during the winter the warm ocean influences its climate, whereas Yakutsk is situated in the interior of the cooled continent of Asia, where the development of surface temperature inversions are characteristic of the winter.

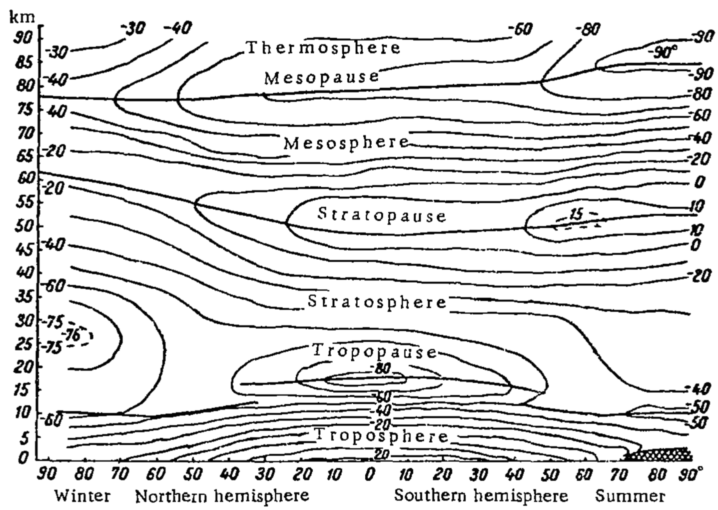

FIGURE 38. Distribution of the mean temperature with height between the surface of the Earth and the 90-km level, January

The thin, solid, and broken lines are isotherms.

A general idea of the temperature distribution with height in the troposphere, the stratosphere, and the mesosphere is given by a vertical cross section of the atmosphere extending from the Equator to the Poles (Figure 38). In the northern hemisphere it is winter, while in the southern hemisphere it is summer. From the isotherms it can easily be seen that the highest temperatures are observed in the southern hemisphere, and the lowest temperatures — at high latitudes of the northern hemisphere.

At high latitudes of the northern and southern hemispheres the tropopause is situated on the average at a height of 9–12 km. Between the Equator and 30° NL and 30° SL it is situated at a height of 16–18 km.

At low latitudes the temperature at the tropopause reaches -80°C, and at middle latitudes -50° to -60°C, since the temperature in the troposphere drops with height, and in the lower stratosphere outside the tropical zone the temperature almost does not vary with height. Thus, in the upper troposphere the lowest temperatures are observed not over the Poles, but over the Equator and the tropics of the northern and southern hemispheres.

At heights of 30–35 km the temperature during the winter is between -73° and -76°C, and during the summer between -30° and -35°C. Higher up the temperature everywhere rises with height up to the level of 50–60 km, where during the summer it exceeds 10°–15°C above zero, and during the winter it is between 0° and -10°C. Still higher, the temperature again begins to drop, and at the level of 80–85 km the temperature during winter reaches -40° to -70°C, and during summer, -70° to -90°C. Above the level of the mesopause, in the thermosphere, the temperature rises continuously with height.

These are, according to modern data, the seasonal variations of the temperature at higher elevations. Let us consider the factors giving rise to such a temperature distribution in the troposphere, stratosphere, and mesosphere.

Actinometric observations show that the effective radiation, i.e., the difference between the downward and upward fluxes of thermal radiation, varies with height both in the troposphere and in the stratosphere. In the troposphere it increases and in the stratosphere it decreases with height. This indicates the decrease with increasing height of the radiative heating in the troposphere and the increase of such heating in the stratosphere. Under such a radiation regime, there should exist between the lower troposphere and the upper stratosphere, i.e., near the tropopause, a layer of radiative equilibrium. This is confirmed by the data given by a number of authors who observed negligibly small radiation heat inflow to the tropopause.

In the stratosphere, the solar radiation is absorbed by ozone and water vapor, and the long-wave terrestrial radiation is absorbed by ozone, water vapor, and carbon dioxide, CO_2. As shown by calculations, the total amount of absorbed and emitted long-wave radiation by water vapor and CO_2 is negative. The radiative cooling of the stratosphere caused by this, is compensated only partially by the absorption of long-wave radiation by the ozone. The total long-wave radiation balance of the stratosphere is therefore negative.

The ultraviolet radiation of the Sun absorbed by ozone, and the infrared thermal radiation absorbed by water vapor, compensate for the long-wave radiation. Calculations show that a basic role in the formation of the temperature field is played by the ozone, which absorbs four times as much heat as water vapor.

Thus, the thermal regime of the stratosphere and mesosphere is determined mainly by the solar radiation and by the capacity of ozone to absorb and to emit the solar ultraviolet radiation.

One sees from Figure 38 that the annual variations in temperature are largest at high latitudes of both hemispheres, whereas at low and even at middle latitudes they are small. This distribution of the annual temperature variations is due to the fact that the air in the polar regions does not receive solar radiation during one half of the year (during the Polar Night).

In addition, during the winter over the Pole, the dark air layer extends to a height of 560 km, and over 80° latitude – to a height of 170 km and more. The heights of the shadowed part of the atmosphere in the cold half of the year are given in Table 23. As can be seen from this table, for about a month of the winter the Pole is in complete darkness up to a height of 500 km.

TABLE 23

Shading heights in the atmosphere

Date	Latitude, degrees					
	90		80		70	
	Sun's angle of declination	Height, km	Sun's angle of declination	Height, km	Sun's angle of declination	Height, km
1 Oct	2°54'	6				
15 Oct	8 14	64				
1 Nov	14 11	197	4°11'	13		
15 Nov	18 17	335	8 17	64		
1 Dec	21 41	472	11 41	130	1°41'	2
15 Dec	23 13	544	13 07	170	3 07	9
22 Dec	23 26	566	13 26	176	3 26	11
1 Jan	23 08	544	13 08	170	3 08	9
15 Jan	21 15	446	11 15	121	1 15	1
1 Feb	17 18	299	7 18	52		
15 Feb	12 15	140	2 15	6		
1 Mar	7 51	57				
15 Mar	2 25	6				

The shading heights in the atmosphere, given in Table 23, were calculated by M. Shabel'nikov for the northern hemisphere. However, in the southern hemisphere too, the shading of the atmosphere in the winter half of the year (October–May) is similar.

For this reason, in the winter half of the year at high latitudes of both hemispheres, the air in the upper stratosphere, owing to the radiation of ozone, cools down (during January to -70°, -75°C in the northern hemisphere, and to -85°, -90°C in the southern hemisphere).

We see from Table 23 that the best conditions for air cooling in the Arctic and in the Antarctic exist in mid-winter. During November and February in the Arctic and during October and August in the Antarctic, the illumination conditions do not contribute to intensive cooling of the air in the stratosphere. Therefore, on the average, the lowest temperatures in the stratosphere are observed during December and the first half of January in the Arctic, and during June and the first half of July in the Antarctic.

During the summer half year at high latitudes, the Sun does not set below the horizon and solar radiation is absorbed continuously by the ozone. This leads to the heating of the air in the stratosphere up to -30°, -35°C, thus causing the large differences between the July and January temperatures at high latitudes.

The small interseasonal temperature variations at low latitudes are due to the small variation in the ozone content and in the intensity of the solar radiation throughout the year.

Although the seasonal temperature variations in the stratosphere correspond to the radiation conditions, sharp temperature rises are sometimes observed in the stratosphere above the Central Arctic during the Polar Night.

The temperature in the troposphere

In order to complete the picture of the temperature distribution in the troposphere, maps showing the differences in the heights of the 300- and 1000-mb isobaric surfaces are given (see pp. 106-107). The differences are proportional to the mean temperature of the air layer between these surfaces, and therefore the isolines on the maps are essentially isotherms. Thus, small values of the isolines indicate cold regions, while large values indicate regions of heat. Since the 300-mb surface is situated at approximately the 9-km level, and the 1000-mb surface is situated close to sea level, the relative-topography maps given in Figures 39 and 40 characterize the mean temperature of an air layer about 9 km thick.

Let us consider some features of the mean temperature distribution in the troposphere during January and July according to these maps. In accordance with the conditions of the inflow of solar energy in the Arctic and in the Antarctic, the air is considerably colder there than at low latitudes, and is independent of the time of the year. Thus the horizontal temperature gradients in the whole troposphere are directed from low to high latitudes, and a wide warm belt extends over the equatorial zone. During the northern winter it is somewhat displaced toward the southern hemisphere, and during the northern summer — toward the northern hemisphere. In addition, the density of the isohypses, characterizing the horizontal temperature gradient, indicates that both in the northern and the southern hemispheres the gradient during winter is larger than during the summer. This can be accounted for by remembering that in both hemispheres the radiation balance in extratropical latitudes is negative during the winter and positive during the summer.

The distribution of the mean January and July temperatures in the lower 9-km atmospheric layer differs considerably from that of the same months near ground level. The highly complex system of isotherms caused by the influence of the underlying surface (see Figures 18 and 19), is in no way reflected on the maps of the relative topography. The influence of the continents and the oceans is clearly visible in the configuration of the temperature isolines even of such a thick air layer. It can be seen from Figure 39 that the January isotherms do not follow the parallels, but are considerably disturbed; this is a result of the influence of the continents and of the oceans on air masses moving from west to east. In fact, a cold trough is situated over the cold continents of Asia and America during

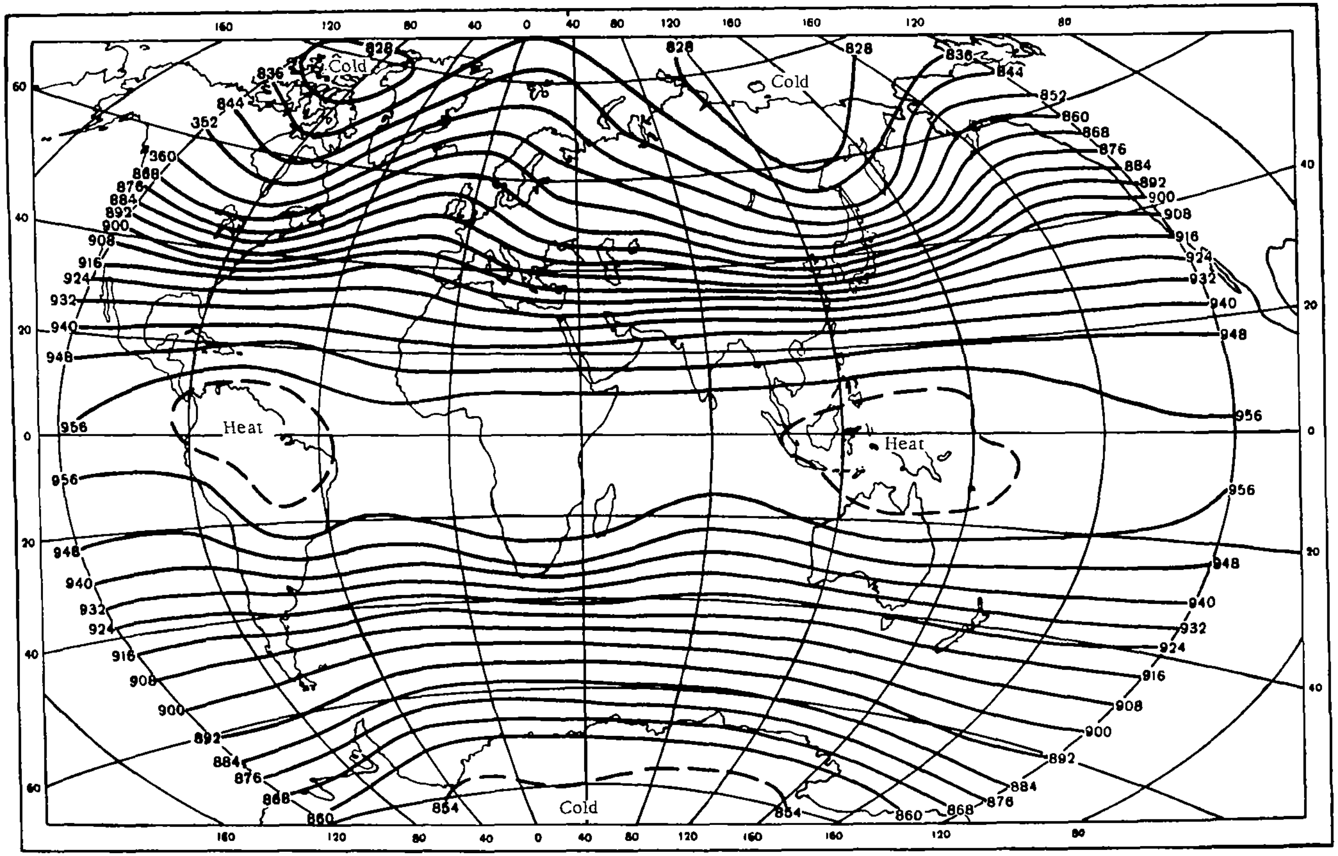

FIGURE 39. Relative-topography map of the 300-mb surface and the 1000-mb surface, January

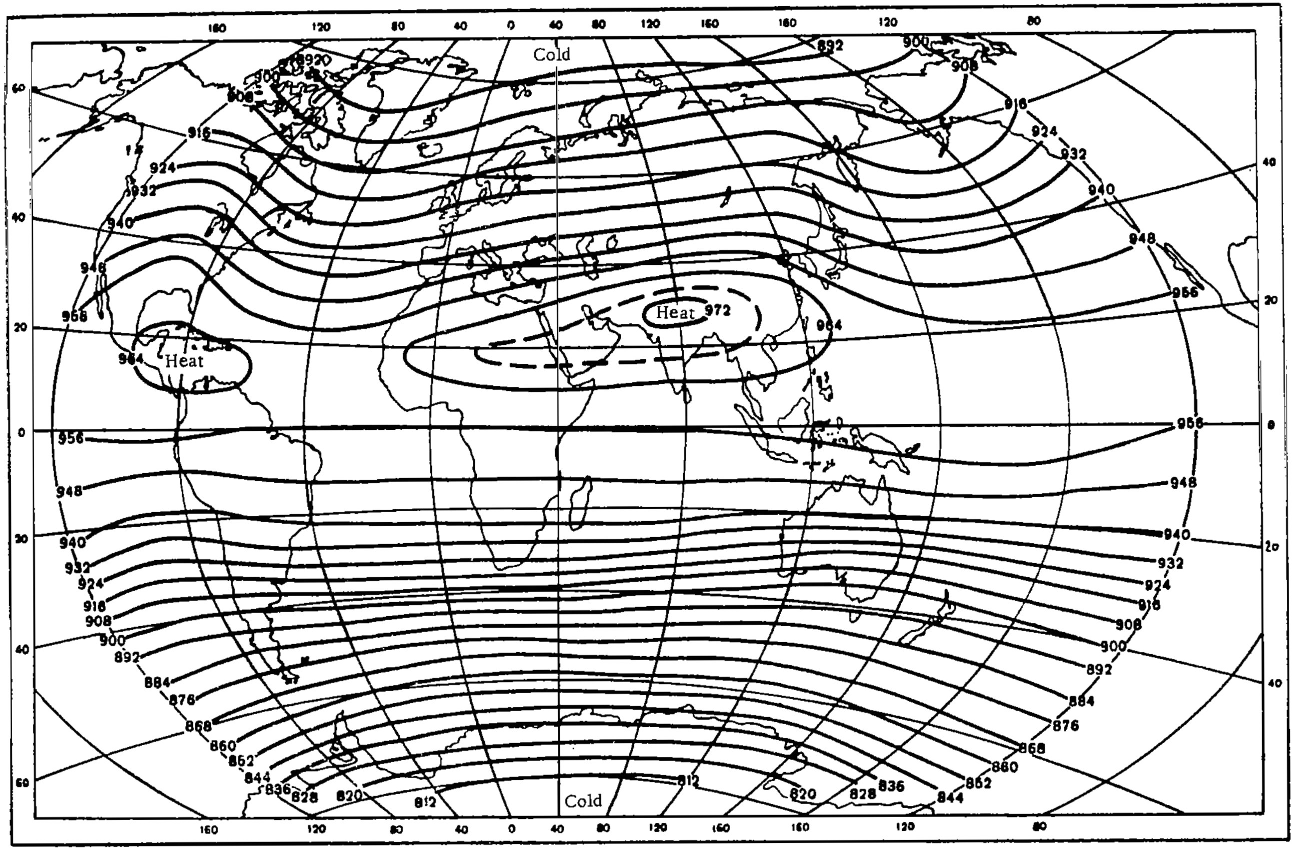

FIGURE 40. Relative-topography map of the 300-mb surface and the 1000-mb surface, July

January, and a warm ridge is situated over the warm oceans. The mean temperature of the layer in the equatorial zone during winter is about 0°C, whereas in the Arctic and in the Antarctic it is -39°C and -30°C, respectively.

The temperatures along the parallels over the continents and over the oceans in the northern hemisphere are equal during July. This is reflected in the configuration of the mean isotherms of the 9-km layer (Figure 40). Here small cold troughs can be observed only over the northern, relatively cold parts of the Atlantic and Pacific oceans. At low latitudes over North America and South Asia, closed heat regions are formed as a result of the intensive heating of the air masses.

The mean temperature of this layer during summer in the tropics exceeds 0°C, and in the Arctic and Antarctic reaches -20°C and -43°C, respectively.

The same features of the temperature field can be observed on the continents of the southern hemisphere, the only difference being that there they are weaker, because of the small dimensions of the continents.

TABLE 24

Mean temperatures of the air layer bounded by the surface of the Earth and the 5-km level for various months, °C

Place, region	Latitude	Longitude	Jan	Feb	Mar	Apr	May	June	July	Aug	Sept	Oct	Nov	Dec
Moscow . . .	55°45' N	37°35' E	-20	-19	-19	-13	-8	-2	4	2	-3	-8	-12	-18
Irkutsk . . .	52 14	104 32 E	-26	-25	-20	-13	-7	0	4	2	-5	-13	-20	-24
Barrow . . .	71 23	156 17 W	-24	-25	-25	-22	-15	-9	-6	-8	-13	-18	-24	-25
Atlantic Ocean . . .	50 00	30 00 W	-6	-6	-6	-4	-3	0	1	1	0	-2	-4	-6
Pacific Ocean . . .	50 00	180 00	-14	-14	-12	-9	-7	-2	0	1	-1	-3	-3	-11
Berlin	52 29	13 23 E	-13	-14	-13	-9	-6	-1	2	2	-1	-4	-9	-13
Vienna . . .	48 13	16 33 E	-13	-14	-13	-10	-5	0	2	3	-1	-3	-8	-13
Agra	27 12	78 00 E	3	3	9	13	15	18	15	15	14	12	8	4

Table 24 gives the mean temperatures of the air layer between ground level and the 5.0–5.5-km level over various places in the northern hemisphere. These data show that in the northern countries the mean temperatures in the lower half of the troposphere remain negative throughout the year (Point Barrow). In Moscow and Irkutsk the temperatures are equal during July and August, although during the winter in Irkutsk it is colder than in Moscow. It is interesting that at the same latitude it is considerably warmer over the Atlantic Ocean during winter than over the Pacific Ocean. The Gulf Stream has a considerable influence over the Atlantic. At low latitudes the mean temperature in the lower half of the troposphere does not drop to negative values for a considerable part of the year.

The temperature in the 9–16-km layer

The temperature regime of the stratosphere differs sharply from that of the troposphere. We have already seen that above the tropopause the temperature may drop or rise or remain unchanged with increasing height.

Maps of the relative topography of higher layers give the general picture of the temperature distribution. We confine ourselves here to only two such maps, which represent the temperature field in the layer between the isobaric 100- and 300-mb surfaces, i.e., between 16 and 9 km for January and July (Figures 41 and 42).

The first factor which determines the difference between these and the preceding maps (Figures 39 and 40) (which also represent the temperature field for the same months) is the unequal isoline density and the variability of the cold and heat focuses. On the maps of December – January in the lower stratosphere, as in the troposphere, a cold focus is situated in the north, caused by the cooling of the air in the ozone layer during the Polar Night.

However, the configuration of the isohypses (isotherms) is different, since in the troposphere the temperature is determined by the heat inflow from the underlying surface (cold continents and warm oceans), and in the stratosphere – mainly by radiation heat exchange. During January (Figure 41) therefore, a cold focus is observed in the center of the Arctic, where the Polar Night dominates. A second wide cold region almost encircles the low latitudes, where the troposphere extends to heights of 16–18 km, and the upper troposphere is the highest, and therefore the coldest (-70°, -80°C). The stratosphere of the middle latitudes of the northern hemisphere is relatively warm, since above the tropopause, at the 10- to 11-km level, the temperature does not vary considerably with height, but remains on the average within -50° to -60°C.

During the northern winter (December-February) an extensive heat region is formed in the stratosphere over the Antarctic due to the heating of the air in the ozone layer during the Polar Day during the southern summer.

From June until August the temperature field in the 9–16-km layer changes sharply (Figure 42). As in the tropospheric case, in the lower stratosphere over high latitudes of the southern hemisphere a cold region is formed caused by the cooling of the air in the ozone layer under the conditions of the Polar Night. In the Arctic at this time of the year, the temperature reaches its highest values, and a cold focus is situated at the equatorial zone from December until February.

A weak heat region is observed in the southern hemisphere between the latitudes 30° and 40°, whose formation is similar to the formation of such a region in the northern hemisphere between December and February. Another small heat focus exists over Central Asia owing to the intensive heating of the air over the desert and mountain ranges.

Relationships between temperature, pressure, and wind fields at heights

The difference between pressure fields at the surface of the Earth and heights is more significant than that between the pressure and temperature fields at heights. This can easily be seen by comparing the respective pressure distribution maps at the surface of the Earth and the maps of the absolute and relative barometric topography.

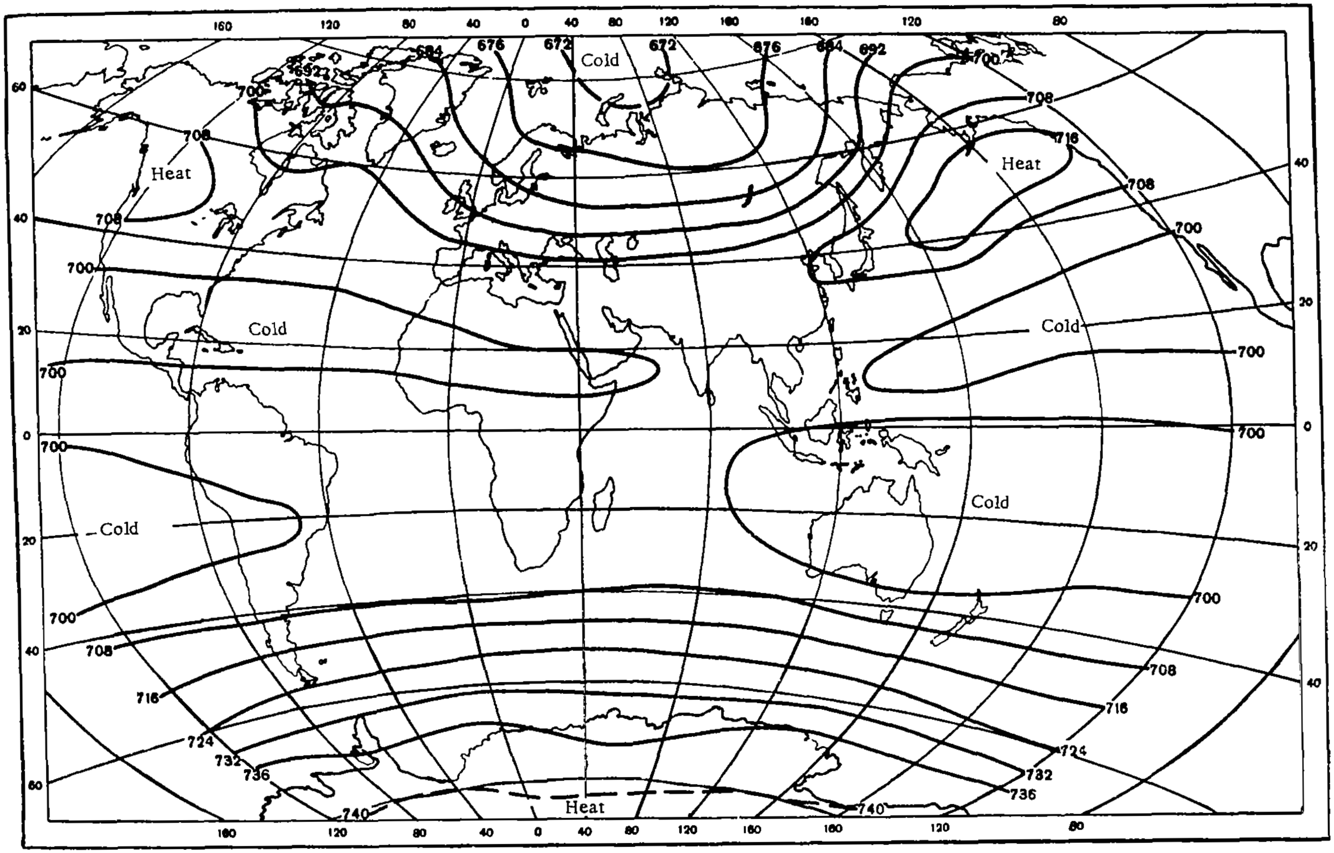

FIGURE 41. Relative-topography map of the 100-mb surface and the 300-mb surface, January

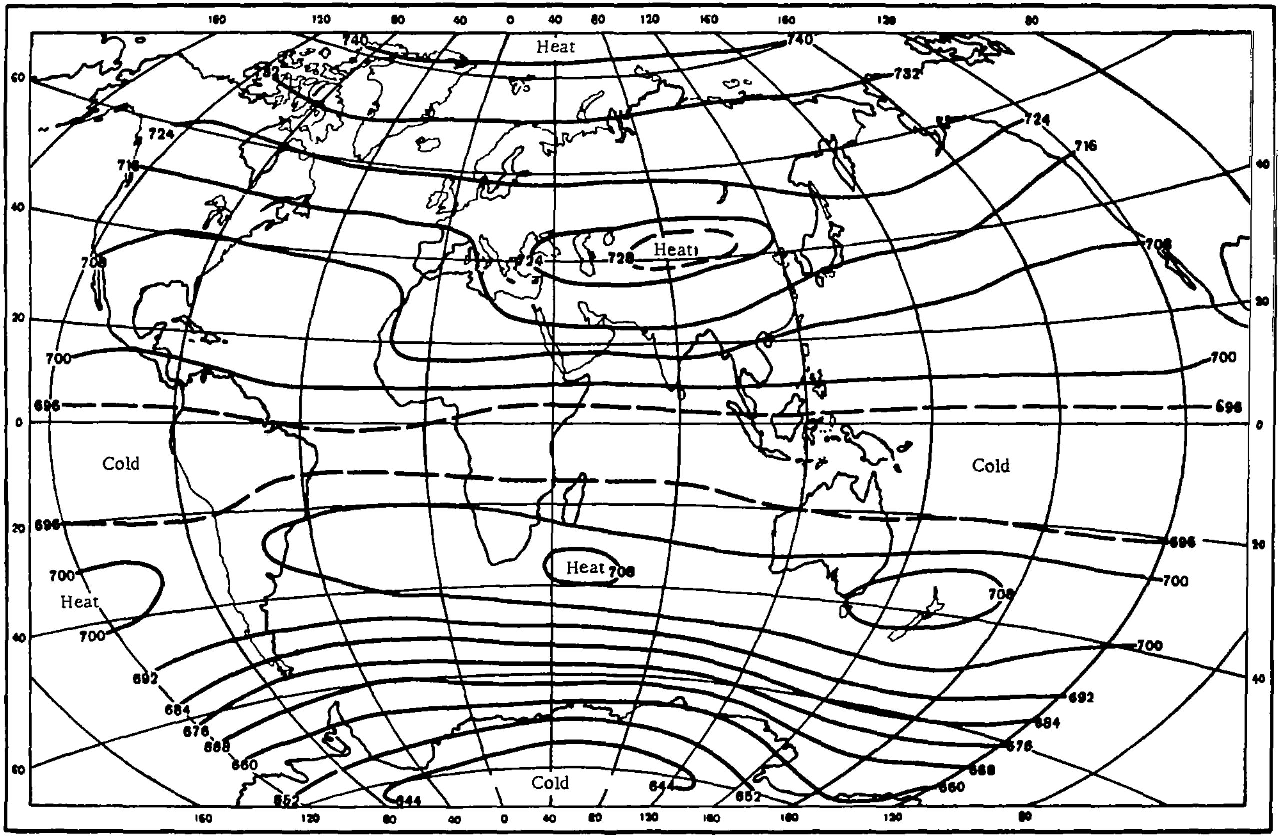

FIGURE 42. Relative-topography map of the 100-mb surface and the 300-mb surface, July

With increasing height, the structure of the pressure field approaches the structure of the temperature field. Already at a height of 5 km the pressure field is more similar to the temperature field than to the pressure field at ground level. Cold regions in the troposphere coincide with low-pressure regions, i.e., with cyclones, and heat regions coincide with high-pressure regions, i.e., with anticyclones. The similarity between the pressure distribution and the temperature distribution at higher elevations also applies to the wind direction and velocity. Since the horizontal temperature gradients usually increase with height, the pressure gradients follow suit, and since the wind velocity depends on the horizontal pressure gradient, it too increases with height. It is interesting to compare the maps of the absolute topography of the 300-mb surface, characterizing the pressure and the air current fields at the 9-km level during winter and summer (see Figures 33 and 34), with the respective maps of the relative topography, characterizing the mean temperature of the thick layer lying below the 300-mb surface (see Figures 39 and 40). Such a comparison shows that not only do the low-pressure regions closely coincide with the heat regions, but even the density of the isolines and the directions of the horizontal temperature and pressure gradients almost coincide. On the other hand, the structure of the pressure field at ground level (see Figures 29 and 30) is totally different and the horizontal gradients are appreciably smaller.

The troughs of the Arctic cold region, penetrating during the winter far to the south, toward the continents, create large horizontal temperature and pressure gradients. To the west of the regions of maximum isotherm density, i.e., over Asia and North America, the isotherms and isohypses converge while to the east, i.e., over the north of the Atlantic and Pacific oceans, they diverge. The divergence of the isotherms and isohypses increases over the oceans due to the fact that the temperature of the surface ocean water is represented by a system of diverging isotherms, and the air temperature field tends to coincide with the temperature field of the ocean surface. As a result of this heat, high-pressure ridges are observed during January over the oceans.

On the January maps of the absolute and relative topography in the northern hemisphere it is possible to observe a zone of highest density of the isohypses and isotherms, almost bordering the northern hemisphere between the latitudes 30° and 50°. In this zone of maximum density of the isobars and isotherms (planetary high-level frontal zones), characterized by maximum temperature contrasts and air-current velocities, active cyclonic and anticyclonic activity with frequent weather changes takes place.

These processes are particularly intense at the eastern shores of Asia and North America, where the mean temperature variation over a distance of 1000 km along the horizontal temperature gradient reaches on the average 13°C during winter and the mean velocities of the air currents in the upper troposphere are about 180–230 km/hr.

The diverging system of isohypses and isotherms over the oceans remains to the north of 40° NL during the summer and at the western shores of the continents even an increase in the convergence is observed. The peculiarities of the heating and cooling of the continents and of the oceans explains this. Since the continents are heated during the summer, heat ridges are formed over the south of North America and over North Africa, thus causing the increased convergence of the isotherms and isohypses there.

Let us discuss this in some detail. By comparing the increase of the mean temperature of an air column 9 km high, one finds that over the Atlantic, at 40° NL, 50° W. long., the mean temperature of the column rises from January to July by only 6°C. At the same latitude over the Balkans the rise is over three times as large, i.e., 20°C. Over the Pacific Ocean at the same latitude and at a longitude of 180° the rise in the mean temperature of such an air column is 10°C, whereas over America at a longitude of 110° W. it is 20°C. These differences in the temperature rise from winter to summer are due to the appearance of heat ridges over the continents and to the decrease in the divergence of the isotherms and isohypses over the oceans during summer.

Similar processes take place in the southern hemisphere, but, as was said above, they are hardly noticeable there due to the relative uniformity of the underlying surface.

Obviously the actual distribution of the temperature in the troposphere deviates from the mean temperature. These deviations are caused by the interlatitudinal transport of warm and cold air masses, brought about by cyclonic and anticyclonic circulation.

Often during the winter over the Arctic one observes a heat region both in the troposphere and the stratosphere, and over middle latitudes of the northern hemisphere – a cold region. This is a result of powerful processes of the interlatitudinal air exchange, when as a result of prolonged and intensive cyclonic activity cold air masses penetrate from the north to the south to several thousands of kilometers and warm air masses are carried in the opposite direction.

The interlatitudinal exchange tends to cause the thermal differences between low and high latitudes to vanish. As we shall see below, owing to this exchange, qualitatively different air masses meet, and zones with large horizontal temperature, humidity, and pressure gradients and large wind velocities are formed.

The interlatitudinal exchange is brought about by cyclones and anticyclones, which are most intense at middle latitudes. The transport of cold air masses from the north to the south usually takes place in the tail section of cyclones and in the front section of anticyclones. Conversely, warm masses are transported from the south to the north in the front of cyclones and in the tail of anticyclones.

The sharp weather changes, whether cooling or warming, which are characteristic mainly of middle latitudes, are connected with the meridional circulation of the atmosphere.

THE WATER CYCLE IN THE ATMOSPHERE

Precipitation

The main source of water vapor in the atmosphere is the huge surface of the oceans, amounting to about 3/4 of the entire surface of the terrestrial globe. The amount of evaporation from the oceans far exceeds that from the land. This can be seen from a comparison of the following figures: According to calculations made by M.I. L'vovich and M.I. Budyko, on the average 519,000 km^3 of water evaporate annually from the entire surface of the globe. Of this amount, 448,000 km^3 comes from the oceans, and only 71,000 km^3 from the land surface. The evaporating water condensing and falling in the form of precipitation is in a continuous cycle.

Most of the water evaporating from the surface of the oceans returns in the form of precipitation, and only a small part is transported by air currents from the oceans to the continent, where it also condenses and falls in the form of precipitation. Precipitation falling on land partially evaporates and clouds and precipitation are formed again. Another part is carried by air currents to the oceans. A continuous exchange of water between the oceans and the continents takes place, the constant level of the oceans being maintained by the inflow from rivers.

The water contained in the air moves with the air both horizontally and vertically. The vertical movements are those which cause the moisture redistribution in the troposphere.

Before clarifying the role of the circulation in the distribution of precipitation, let us consider the mean annual amounts of precipitation at various latitudes in both hemispheres, obtained on the basis of distribution maps of the annual amount of precipitation over the terrestrial globe, compiled by O.A. Drosdov (Table 25).

It follows from the data given in Table 25 that the annual amounts of precipitation are distributed nonuniformly with latitude. Thus, for example, the largest amount of precipitation falls in the equatorial zone, the smallest amount — at high latitudes. The second maximum of the mean annual amounts of precipitation, both in the northern and in the southern hemispheres, is observed between the latitudes of 40° and 60°. At middle latitudes of the northern hemisphere the amount of precipitation is somewhat larger than at high latitudes.*

The mean amount of precipitation is equal to 720 mm, according to the calculations of M.I. L'vovich. However, this mean value hides a great variation in the distribution of the precipitation.

* There is still insufficient data on the amount of precipitation at high latitudes of the southern hemisphere.

The annual, seasonal, and monthly amounts of precipitation do not only differ at points situated a considerable distance apart. They may differ at points very close to each other. This often occurs in mountains, where even at a distance of less than 100 km, the amounts of precipitation in different months or seasons differs sharply. The variation in the distribution of the precipitation can be explained.

TABLE 25

Distribution of the mean amount of precipitation at various latitudes, mm

Latitude, degrees							
80–70	70–60	60–50	50–40	40–30	30–20	20–10	10–0
Northern hemisphere							
194	340	510	561	501	513	763	1677
Southern hemisphere							
–	–	976	868	564	607	1110	1872

The distribution of the precipitation over the continents is determined by a variety of factors. Chief among them is the continual transportation of water vapor by the air.

How the clouds are formed

We are already familiar with a remarkable property of water vapor which distinguishes it from the other gases of the atmosphere, i.e., the variation in the amount of vapor per unit air volume with the temperature. The variations in the amount of water vapor contained in air are quite considerable. Instead of unit volume consider unit weight of air. We find that at a temperature of 27°C, for example, 1 kg of air may contain a maximum of 23 grams of water vapor, and at 0°C – only 4 grams. At low temperatures the amount of water vapor in air is negligibly small. For example, 1 kg of air at a temperature of 33°C below zero may contain only 0.2 grams of water vapor. This is 115 times less than that contained in 1 kg of air at a temperature of 27°C above zero. Therefore at middle and high latitudes the amount of water vapor rapidly decreases with increasing height and a half of all the vapor contained in the troposphere is concentrated in the layer from the surface of the Earth to a height of 1.5 km.

Many processes which are so important for the existence of life on Earth e.g., condensation, formation of various forms of clouds, precipitation, as well as evaporation, are connected with this remarkable property of water vapor.

As we know, air becomes saturated with water vapor when the amount of the latter at a given temperature reaches a maximum. Therefore, if saturated air is cooled, excesses of water vapor appear and condense, i.e., pass to the liquid or solid state, and fall in the form of precipitation. The character of the precipitation (liquid or solid) depends on the air temperature. If saturated air is heated, the condensation stops. Favorable conditions are then created for evaporation from the surface of seas and oceans, from the wet surface of the Earth, from vegetation, and from every place where there are water reserves; the air, tending to replenish the moisture deficiency, absorbs the missing amount of heat at the given temperature. Under favorable conditions the air is enriched by vapor even by evaporation from snow cover and from glaciers.

The further the air from the state of saturation, the faster the evaporation. Thus on clear days, the evaporation from the humid surface of the Earth and from the surface of water basins is most intensive. At night, when the air is cooled and approaches the state of saturation, the water vapor condenses, leading to the formation of fog, and dew. In these cases the evaporation from the surface of the Earth stops.

In order that condensation of water vapor leading to the formation of clouds will take place, an excess of water vapor above saturation is necessary. Such an excess may appear either as a result of an increase in the vapor content of the air, or as a result of a drop in its temperature below the dew point.*

An increase in the moisture content of the air occurs as a result of evaporation from the underlying surface. The air temperature drops either as a result of its contact with the cold underlying surface and of radiation, or when the air ascends, expands, and adiabatically cools down. In nature, both factors are usually effective, but often large volumes of air are cooled in upward motion. The increase in the moisture content due to evaporation is slow, and only rarely has a decisive effect.

The air is cooled significantly as a result of nocturnal radiation from the Earth and the tops of the clouds. The radiation intensity from the Earth and, consequently, the air cooling, depend on the cloud coverage. Particularly intense cooling occurs in the surface air layer in cloudless weather; this often leads to the formation of fog. Nocturnal radiation at the tops of the clouds gives rise to an increase in their vertical thickness and water content. But the main reason for the formation of clouds, as already stated, is the adiabatic expansion taking place in ascending air motion.

Several reasons for vertical air motion are known. The most important one is the variation of the air currents with time. Thus redistribution of the air masses, associated with large-scale vertical motion, takes place. Quantitative calculations of the vertical velocity of air motion were made by A.F. Dyubuk, A.S. Zverev, G.I. Morskii and others. The vertical velocity is low, on the average about 3–5 cm/sec. However, if we take into account that the ascent or decent of air masses takes place over a long period, the immense role played by the ascending motion of large air volumes in the formation of clouds and precipitation becomes clear. In fact, assuming that the mean velocity of ascent of the air is 4 cm/sec, then an air mass may rise by more than 3 km during 24 hours and under usual conditions may cool down by 20°–30°C. For average moisture content of the air, such cooling is sufficient for the formation of a thick cloud layer and the fall of much precipitation.

* Dew point is the term applied to the temperature at which the air becomes saturated.

When the air stratification is unstable, large air masses are also made to rise by thermal convection. In this case the velocity of ascent often attains 10 m/sec and more, and therefore the formation of convective clouds and precipitation is most intense.

Other reasons for vertical motion are air friction with the surface of the Earth, turbulence, the encounter of an air current with mountainous obstacles, etc.

In some cases (in particular in cyclones) the friction causes convergence of the currents and ascending air motion, in others (particularly in anticyclones) it causes divergence of the currents and descending air motion.

When encountering mountain ranges, the air tends to flow around them. However, if the obstacle is of considerable width, the air is lifted along the slopes and crosses the crest to the lee side. When the air stratification is unstable, its ascent along the windward slopes of the range is stormy. Therefore, in a stably stratified air mass, stratified clouds are formed on the windward side of mountain ranges, and continuous precipitation of low and moderate intensity falls. This is most often observed during winter. During the summer, air masses with unstable stratification, upon encountering obstacles, turn upward with high vertical velocities, which leads to the formation of thick cumulus and cumulonimbus clouds, resulting in torrential precipitation.

The formation of clouds and precipitation seems very simple at first. As a result of the ascent and cooling of the air, the water vapor should condense, and water droplets, coalescing with one another, get larger and fall in the form of precipitation. However, the formation of clouds and precipitation is actually a very complex physical process. During the last twenty to thirty years, the study of the processes of cloud formation has been conducted not only under laboratory conditions, where in special chambers clouds are created and dispersed by artificial means, but also in their natural environment by observers who, with instruments, fly in laboratory-airplanes.

Many details of the nature of cloud formation became better understood recently. This does not mean that before our time scientists did not have any idea about the physics of clouds. Already in the middle of the 18th century M.V. Lomonosov, studying the nature of atmospheric electricity, related charged cloud water drops to ascending air motion. Later on, a series of reasonably correct theories as to the conditions for the formation of large water droplets in clouds were postulated both in Russia and abroad.

Let us sketch briefly the position of investigations today.

Clouds are formed mainly in the troposphere. They differ by their structure, form, and height. Accordingly, there are small and large droplets, liquid and solid precipitation, and so on. In order to understand the formation of the various kinds of clouds and atmospheric precipitation, it is necessary to know the microphysical features of their structure, and particularly of their phase structure (i.e., whether they consist of water droplets or of ice crystals), the water content, the reason for droplet growth, and so on.

Before studying the microphysics of clouds, we will classify them.

Classification of clouds

According to the phase structure, there are water, ice, and mixed clouds. According to international classification, clouds are divided with respect to their height into three levels: low, middle, and high clouds. As to form, they are divided into the following basic groups: stratus clouds, stratocumulus clouds, and cumulus clouds. Stratus and stratocumulus clouds are formed during a slow ascent and adiabatic cooling of the air. Cumulus clouds result from a rapid vertical ascent of the air and usually extend to the upper troposphere.

According to the existing classification, to the upper-level clouds belong high cirrus, cumulocirrus, and stratocirrus clouds; to the middle-level clouds – high-cumulus and high-stratus clouds; to the low-level clouds – stratocumulus, and stratus clouds; there are also the thick cumulus and cumulonimbus clouds.

Upper-level clouds consist of ice crystals. They are distinguished by a streaked structure. Cirrus clouds often appear as the result of a disintegration of the upper part of cumulonimbus clouds, forming an "anvil".

High-stratus clouds often form thick layers, from which, depending on the air temperature, rain or snow falls.

Stratus clouds of the lower- and middle-levels are water or mixed clouds. Their densities are different. When their density is low the Sun may be seen through them. If the stratus clouds consist of ice crystals, then light circles (halo), resulting from the refraction and reflection of light in the ice crystals, are observed about the Sun.

Stratocumulus and high-cumulus clouds are formed when ascending air encounters a layer with temperature inversion, which hinders the further ascent of the air. These clouds are usually not very thick. In the case of a high moisture content in the air and convection in the air layer below the clouds, stratocumulus clouds may give light precipitation.

Cumulus and cumulonimbus clouds are formed as a result of thermal convection. Their vertical thickness depends on the thickness of the unstably stratified air layer. When the instability is small or the air is dry, so-called good-weather cumulus clouds appear. Their vertical development is usually hindered by stably stratified higher lying air layers (above 2–3 km). Conversely, powerful development of cumulus clouds and their transition into cumulonimbus clouds takes place under considerable air instability, extending to the middle and upper troposphere.

For clouds to appear, the air, besides ascending, must contain enough water vapor for condensation to begin when the air cools by several degrees (approximately by 5°–10°C). The larger the moisture content of the air at a given temperature, the lower the condensation level. During the winter it is usually nearer to the surface of the Earth than during the summer.

Condensation of water vapor near the surface of the Earth leads to the formation of fog. The relative humidity usually approaches 100% in this case.

During a fog the condensation level lies at the surface of the Earth.

Water, or liquid-drop clouds consist of water droplets; below the zero degree level the water droplets have a positive temperature, and above this level – a negative temperature, i.e., they are supercooled. Very small water droplets may exist at temperatures of -10°, -20°C, and even -30°C. In the range of from 0°–30°C below zero ice crystals are encountered as often as supercooled drops. At a temperature of -30°C clouds consist, as a rule, of ice crystals. Mixed clouds consist of supercooled water droplets and ice crystals. It has been shown by investigations that in the middle of Europe pure water, pure ice, and mixed clouds are encountered at almost the same frequency. It is natural that pure water clouds most often exist in the warm half year and ice clouds – in the cold half year.

New data on the microphysics of clouds

During recent decades the theory of cloud and precipitation formation has been developed by T. Bergeron and V. Findeizen. It had been assumed that pure water clouds may yield only small amounts of precipitation so that intensive precipitation results from mixed clouds. In addition, it was considered that the solid phase appears when the clouds reach the so-called level of ice crystals, and that in developing mixed clouds the ice particles grow mainly by sublimation of water vapor.

Latest investigations, carried out both in the Soviet Union and abroad, widened the knowledge on cloud structure, on the physics of cloud formation, and on the role of factors affecting the growth of cloud drops and crystals. They established regularities of variation in the size of cloud drops and laid the basis for a quantitative theory of precipitation.

In the Soviet Union systematic studies of the structural features of clouds were begun in the Caucasus Mountains in 1935–1937 under the direction of V.N. Obolenskii. The investigations were extended after the second world war. E.S. Selezneva, V.A. Zaitsev and others studied the sizes of cloud drops and the amount of water in cumulus clouds. It was established that clouds consist of drops with radii of 2 to 70 microns. According to the data of A.M. Borovikov, the radius of drops in stratus and high-stratus clouds is on the average about 7 microns, and in stratus-rainy clouds about 9 or 10 microns.

The velocity of fall of small water drops (up to 10 microns in radius) is negligibly small, i.e., 1 cm/sec; thus, such drops are almost in a suspended state. Larger drops fall faster. In particular, drops with a radius of about 1 mm fall with a velocity of 2 m/sec. Curiously, in stratus clouds the smallest drops are situated mainly at the base of the clouds, and the large drops — near its top. According to data accumulated over many years of observations by V.A. Zaitsev, E.S. Selezneva, and I.I. Chestnaya, the smallest drops are also concentrated at the base of cumulus clouds, and the large drops – in their central and upper portions.

Since the size of cloud drops is very small, several hundreds of them exist per cm^3 of a water cloud. The number of ice crystals in 1 cm^3 of an ice cloud is much smaller; for example, a few cubic centimeters of a cirrus cloud contain one ice crystal or one snow flake.

It is very difficult to determine the number of drops per unit cloud volume. Therefore, the number of drops is often determined from the so-called l i q u i d w a t e r c o n t e n t o f t h e c l o u d, i.e., from the water content in water-drop or solid form per unit cloud volume. The liquid water content is expressed in grams per cubic meter of the cloud. Measurements of the amount of water clouds, made between 1948 and 1950 by A.Kh. Khrgian and V.A. Zaitsev, showed that in clouds the water-drop moisture is as a rule less than the gaseous moisture. It was found that the amount of water in clouds varies within wide limits, the amount of water in water clouds being considerably larger than in ice clouds. Thus, for example, in water clouds 1 m^3 contains 0.3–4.0 gram of water-drop moisture, whereas ice clouds contain 0.1–0.5 gram. This is understandable in view of the fact that ice clouds develop at low temperatures where the air contains a small amount of water vapor.

Investigations showed that for the formation of the primary drops, saturation is insufficient and considerable supersaturation, which in natural conditions is not observed, is required. The process of nucleation or the formation of the primary drops, is facilitated in natural conditions by the fact that water vapor settles on condensation nuclei possessing a hygroscopic nature, i.e., the property to absorb molecules of water vapor. Condensation nuclei are very small particles of hygroscopic substances, soluble in water, on which the water vapor molecules are deposited, e.g., particles of sea salt, or sulfur dioxide, dust and smoke particles, microorganisms, etc. In the atmosphere they exist in large quantities. In the absence of these nuclei formation of droplets would occur under a high supersaturation.

There are more condensation nuclei over industrial towns. According to measurements by I.I. Gaivoronskii at Dolgoprudnyi, a distance of 19 km from Moscow, 1 cm^3 contains up to 20,000 of these nuclei. Experiments and theoretical investigations by B.V. Kiryukhin (1950) and L.G. Kachurin (1951) showed that as a result of the primary condensation, low temperature drops are formed, freeze and become seeds of ice crystals and snowflakes. The rate of formation of ice-crystal seeds depends on the size of the drops, on the air temperature, and on the cooling rate. Crystallization of clouds usually occurs at a temperature of 10°–15°C below zero. Similar results were obtained in 1944 by L. Krystanov, and in 1947–1950 by N.S. Shishkin and by others.

A valuable contribution to understanding phase transformations in clouds was made by V.N. Obolenskii, E.G. Sak, E.K. Fedorov, L.T. Matveev, V.Ya. Nikandrov, Ni. Vul'fson and other scientists.

Seeded drops in clouds grow under the effect of two factors: condensation and coagulation, i.e., drop fusion.

Drop fusion plays an important role in their enlargement, and occurs both upon their collision, and in fall. Methods of calculating drop growth in fall were made by L. Krystanov and N.S. Shishkin. Investigation of the process of drop fusion was carried out also by M.A. Aganin, B.V. Deryagin and others.

The scheme of cloud and precipitation formation can be represented as follows. Due mainly to ascending motion and adiabatic expansion, the air cools and gets saturated. As a result of further ascent and cooling, supersaturation occurs and primary drops are formed. They grow as a result of fusion, the larger the drops the faster their growth. Large drops begin to fall. Since the velocity of a drop depends on its size, insufficiently large drops, having a small drop velocity, are balanced by the ascending air motion, or are lifted to the upper part of the cloud. However, as its size increases, the velocity of the drop increases, exceeding the velocity of ascent of the air. Then large drops fall in the form of precipitation.

For precipitation to fall from water clouds the clouds must have a vertical thickness of over 3 km and a water content exceeding 1 g/cm^3. At high and middle latitudes, such clouds will usually give light or continuous precipitation of intermediate intensity. At low latitudes, where the moisture content of the air is high and the level of the zero isotherm is situated at heights of 5–6 km, water clouds forming as a result of convection have a large vertical thickness and give heavy rain.

Ice clouds usually yield precipitation of intermediate and low intensity, since their liquid water content is lower than that of water clouds.

Most intensive precipitation falls from mixed clouds. This is due to the fact that with the appearance of a solid phase in a medium of supercooled drops, the ice particles have an accelerated growth and the drops rapidly acquire a large size and fall as precipitation.

A quantitative account of the factors affecting the formation of clouds and precipitation made it possible to lay the bases for a quantitative theory of precipitation.*

Atmospheric circulation and the water cycle

The precipitation regime under various physicogeographic conditions is determined to a large extent by the water cycle in the atmosphere. Since moisture is transported together with air currents, the water cycle is also closely related to the atmospheric circulation.

New ways of observing the wind, humidity, and temperature at heights, independently of the weather conditions, gave great possibilities for a deeper investigation into atmospheric processes, including the complex process of the water cycle. It thus became possible to calculate the water cycle over large and small areas and to determine the role of both the moisture transport from the oceans, and of the local evaporation in the formation and fall of precipitation. This is undoubtedly of great importance for the projects carried out in the USSR on the transformation of nature in the arid regions of the country.

For a correct calculation of the water cycle in the atmosphere, we must find out the role of the most important factors taking part in it: moisture transport in the atmosphere, precipitation, runoff, and evaporation. It is particularly important to determine the amount of moisture transported in the atmosphere per unit time.

The water cycle scheme

The precipitation, runoff, and evaporation for average annual conditions are related by:

$$P = E + R,$$

* More detailed information on the physics of clouds and precipitation can be found in the interesting booklet: Kiryukin, B.V. and P.N. Krasikov. Oblaka, dozhd' i sneg (Clouds, Rain, and Snow). —Leningrad, Gidrometeoizdat. 1953; and Shishkin, N.S. Oblaka, osadki i grozovoe elektrichestvo (Clouds, Precipitation, and Thunderstorm Electricity). —Leningrad, GITTL. 1954.

where P is the amount of precipitation falling on a given territory during a given time interval, E, the amount of moisture evaporating from the territory, and R, the river runoff during this time interval.

The amount of precipitation P falling on a given territory is derived partially by the water vapor transported to this territory from the outside (this is the so-called external precipitation P_e) and partially from the water vapor from local evaporation from the underlying surface (this is internal, or local precipitation P_l) (Figure 43). Since the air and its water vapor is in continuous horizontal motion with a velocity quite high as compared with the velocity of the vertical motion, it is natural that the precipitation falling on a small area will consist mainly of moisture transported to this area from the outside. In this case, the role of the local evaporation in the formation of precipitation on the same territory is so small that it can be ignored. The role of the local evaporation in the formation of precipitation falling on a given territory will increase with increasing dimensions of the territory under consideration.

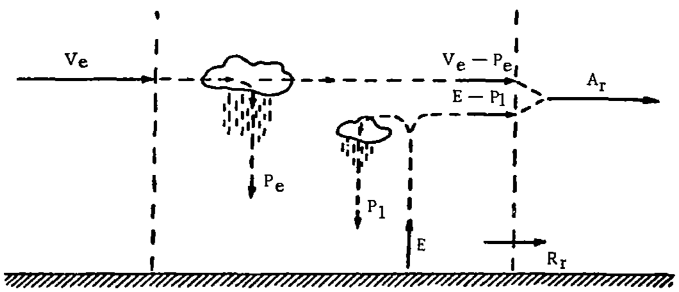

FIGURE 43. The water cycle scheme

When investigating the water cycle in the atmosphere, the importance of the local evaporation in the formation of precipitation is determined. The water vapor V_e arriving from the outside is partially lost in the formation of external precipitation P_e falling on the given territory, and the remaining part $(V_e - P_e)$ is transported beyond its boundaries (Figure 43). This is the atmospheric flow $V_e - P_l$.

The moisture, E, evaporated from the given territory, partially goes to the formation of internal or local precipitation, P_l, and the remaining part is carried beyond the boundaries of this territory. This is also atmospheric flow $E - P_l$. Thus the total atmospheric flow A_r from a bounded territory, consists of the moisture transported from the outside $(V_e - P_e)$ and the moisture evaporating from this territory $(E - P_l)$.

It is very interesting to determine the coefficient of the water cycle, showing how many times moisture arriving from the outside and falling on a given territory in the form of external precipitation P_e, will evaporate and fall again on the same territory until it is carried away by the atmospheric, A_r, and river, R_r, flows. However, of the three quantities involved in the equation $P_e = A_r + R_r$, necessary for the determination of the coefficient of the water cycle, only the magnitude of the river flow, R_r, is measurable. The other two quantities, P_e and A_r, are not amenable to direct measurements. Therefore, until recently, arbitrary assumptions have been made in order to determine the coefficient of the water cycle. Thus, for example, in an attempt to calculate the water cycle over West and Central Europe it was assumed that the magnitude of the atmospheric flow is equal to half the river flow. It was also assumed that summer precipitation in Europe is formed mainly by the water vapor entering the atmosphere as a result of local evaporation.

Recent investigations show that over West and Central Europe, as well as over other regions of middle and high latitudes, the air is in continuous motion and that individual volumes move with high velocities. In particular, according to data of aerological observations, the mean wind velocities over Europe at a height of 3 km are over 30 km/hr. Accordingly, if we take into account the dimensions of the territory of West and Central Europe an air volume moving in one direction may traverse this territory in approximately two days.

The precipitation falling during summer in Europe is connected mainly with the activity of cyclones moving for the most part from west to east. The heating of the air masses arriving over Europe from the Atlantic and causing the development of thermal convection, intensifies the formation of clouds and precipitation from moisture brought there by air masses. The widespread theory that for West and Central Europe the amount of external precipitation is only 1/2 of the total amount of precipitation was thus proved wrong.

However, evaporation from the land surface plays an important role in moistening the air. Already in the 19th century, A.I. Voeikov, concerned with the problem of the water cycle, correctly pointed to the role of evaporation from the continents in the formation of precipitation. He also suggested that precipitation may result from the water vapor transported by the atmospheric circulation to tremendous distances. But at that time the data on the atmospheric circulation and on the amount of transported moisture were completely insufficient for the calculation of the water cycle.

We now have enough data of aerological observations over huge territories of continents and oceans, as well as data on large- and small-scale atmospheric circulation, to be able to solve the problem of the water cycle in the atmosphere. A reliable calculation of the water cycle both over a bounded land territory, and over continents as a whole, can thus be carried out.

Precipitation and the atmospheric circulation

The distribution of precipitation over the continents depends on a number of factors. Chief among them is the atmospheric circulation, which must therefore be studied first. As is shown by the precipitation map, the annual amounts of precipitation over the extensive territory of the Eurasian Continent decrease as one gets farther from the shores of the Atlantic Ocean from west to east (Figure 44). The considerable deviations from this precipitation distribution, observed in certain regions, are due to the presence of mountain regions.

The amount of precipitation is determined not only by the horizontal moisture transfer, but depends to a large extent, as was already said,

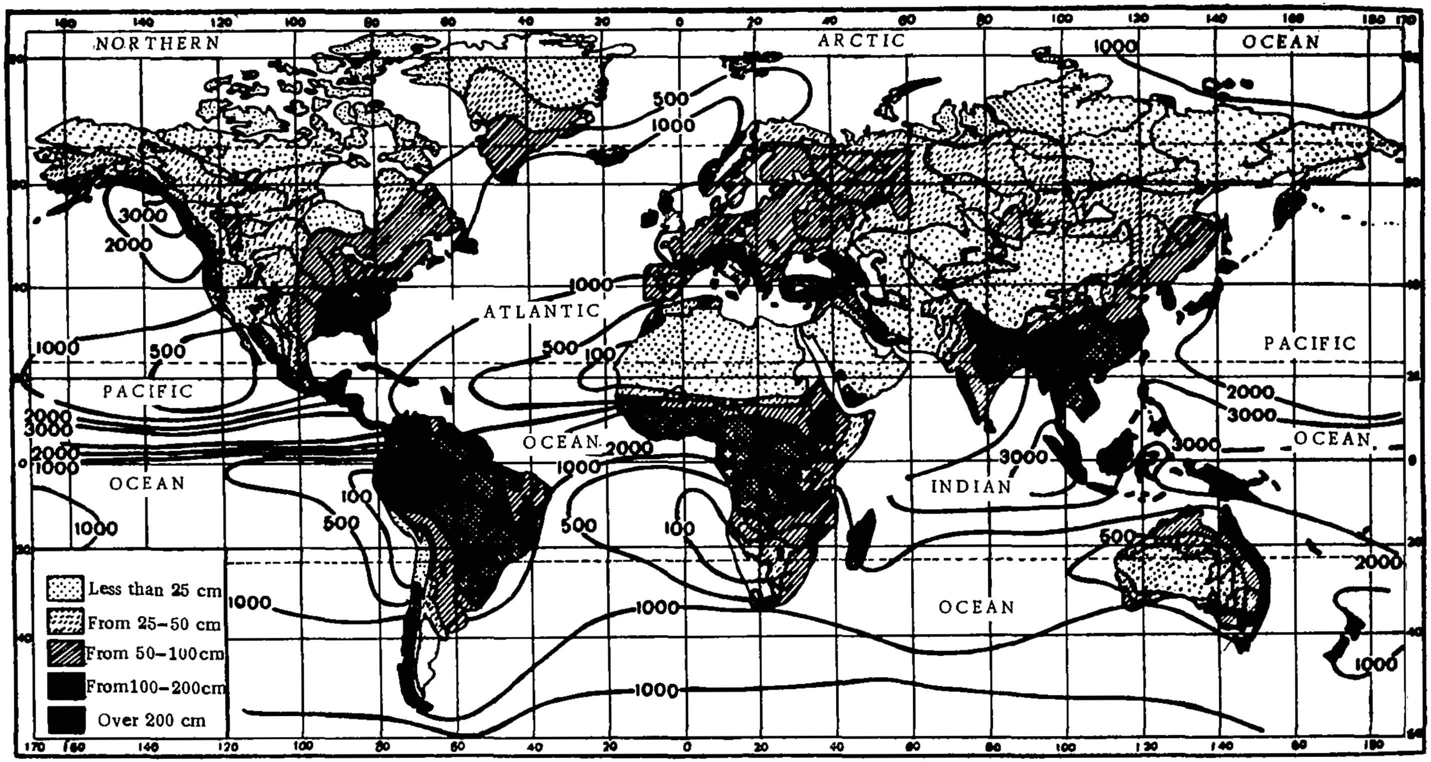

FIGURE 44. Annual amounts of precipitation

on the vertical motion of large air masses containing water vapor. In the system of cyclones, ascending air motion prevails. The air is thus adiabatically cooled, becoming saturated. In the system of anticyclones, on the other hand, descending motion prevails and the air is adiabatically heated and brought further from the saturation state. Therefore, in those regions where cyclonic activity prevails, there is more precipitation than at regions where anticyclonic activity prevails.

It should be noted that precipitation may form not only in the case of cyclonic circulation but also in the case of unstable air stratification, where, owing to thermal convection, air masses move upward with high velocities. In these cases the upward velocity of air containing moisture may reach 5–10 m/sec and more, which is hundreds of times greater than the velocity of ascent of air in cyclones in the absence of convection.

Thermal convection arises upon the heating of air layers adjacent to the surface of the Earth. However, precipitation is not formed in all ascending air currents. In arid regions in the south and southeast of the USSR, intensive ascending currents very often appear during the summer due to the strong heating of the land and its adjacent air layers, but clouds and precipitation are not formed. This is due to the fact that at a high air temperature its moisture content is insufficient to reach the saturation state. Intensive heating of the lower air layers in the absence of evaporation sources leads to an even greater decrease in the relative humidity.

Precipitation at low latitudes

The amount of precipitation is particularly large in those regions where water surface prevails and ascending currents arise. Let us take as an example the equatorial zone, where land constitutes only about 20% of the area. In accordance with the positions of the high-pressure regions in the tropics and subtropics, the trade winds (northeasterly in the northern hemisphere and southeasterly in the southern hemisphere) are directed to the high-pressure equatorial zone. Relatively dry air masses in the trade-wind current, arriving in the calm equatorial zone, get moistened. As a result of the large inflow of solar energy during the year and of the heating of the air, conditions are created for thermal convection, and humid-air currents moving upward form cumulonimbus clouds of large vertical thickness. Over the land the precipitation falls mainly in the second half of the day, and on the oceans it falls at night in the form of tropical downpours, often associated with heavy thunderstorms. Owing to the high moisture content of the air and the large losses of heat on evaporation from the humid soil, the daytime heating of the air is weakened. In addition, the high content of water vapor prevents a strong night cooling. Therefore even on the continents the daily temperature amplitude is usually 8°–12°C, rarely exceeding 14°–15°C. At night even under a small temperature drop, fog often appears.

The amount of precipitation has a well-pronounced annual variation. The prevailing annual amount of precipitation in the narrow equatorial zone amounts to 1500–2000 mm. In the same zone on the eastern coast of Africa, however, there is less than 1000 mm of precipitation, and on the islands of Borneo and Sumatra, in the headwaters of the Amazon river, and at other places – up to 3000 mm per year and more. The greatest amount of precipitation usually falls there from December until March, the smallest amount from June until October. Examples of the distribution of the annual precipitation are given in Table 26, and for a number of other places they are represented graphically in Figure 45. These graphs give, in addition to the annual variation of precipitation, the annual variation of the temperature.

TABLE 26

Annual variation of precipitation in Manaos and Port Darwin, mm

Place	Latitude	Longitude	Jan	Feb	Mar	Apr	May	June	July	Aug	Sept	Oct	Nov	Dec	Total
Manaos	3°07'S	60°02' W	234	228	243	217	179	92	55	35	52	105	139	196	1775
Port Darwin	12 38	130 50 E	400	340	260	110	20	4	2	2	10	50	120	260	1578

However, as was already said, the annual precipitation in the equatorial zone is not the same everywhere. As can be seen from the graphs, on the Pacific Ocean shore of South America at Buenaventura, situated somewhat to the north of the Equator (3°57' NL), the mean annual precipitation is 7129 mm. Somewhat to the south of the Equator, however, at Guayaquil, the annual amount is 976 mm and in Callao (12°SL) – only 25 mm. Such a difference exists also in other regions of the equatorial zone. Thus, for example, in Accra (Ghana) and in Debundscha (Cameroons) situated almost at the same latitude (Central Africa), 686 and 9498 mm of precipitation falls respectively, in Zanzibar (East Africa) – 1596 mm, and in Aden (South Arabia) – only 58 mm.

Almost as much nonuniformity of precipitation is exhibited over the oceans. In Colombo (Ceylon) the annual amount exceeds 2200 mm, in Kikori (New Guinea) — 5850 mm, in Atuona (Marquesas Islands) — 1900 mm, etc.

As can be seen from the precipitation map (see Figure 44), to the north and to the south of the Equator the amount of precipitation decreases. Between the latitudes 20° and 40° in the northern and southern hemispheres (Table 26), the amount of precipitation is considerably smaller on the average. At these latitude zones anticyclonic conditions prevail, so that the annual amount of precipitation over extensive territories does not exceed 100 mm. At the same time, between the latitudes 20° and 30°, in particular in southeast Asia, the amount of precipitation, falling mainly during the summer half year, reaches 1000–2000 mm. On the southern slopes of the highest mountainous region in the world, the Himalayas, the amount of precipitation is over 3000–4000 mm, and in individual places it exceeds 10,000 mm, which amounts to a water column over 10 m high. Precipitation in southeast Asia is caused by trade-wind circulation. There during the summer, humid air currents move from the Indian and Pacific oceans to the low-pressure region whose center is situated over South Asia.

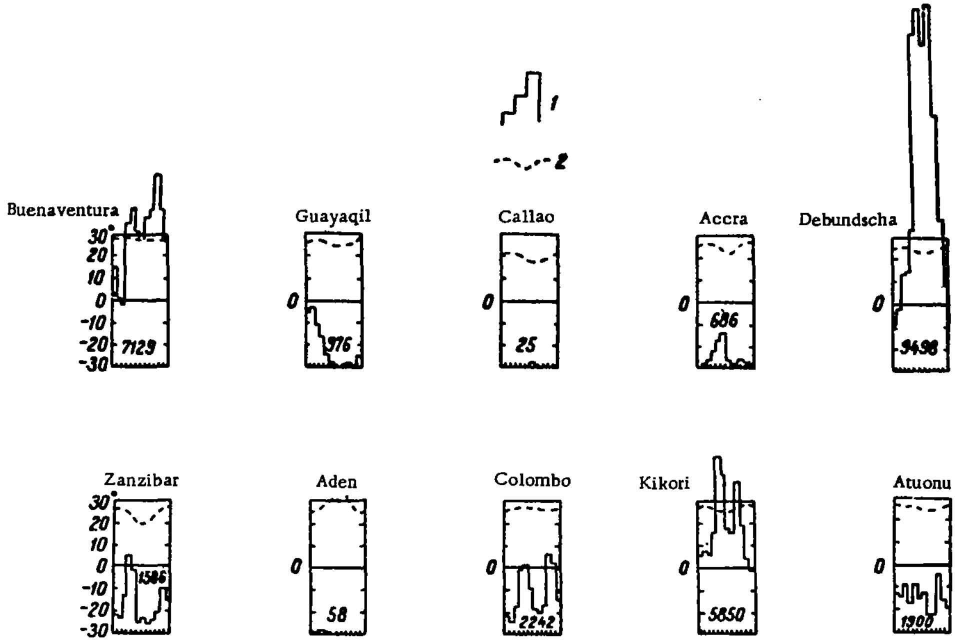

FIGURE 45. The annual variation of the precipitation (1) and temperature (2) at various places in the equatorial zone

The influence of mountain ranges on precipitation

Mountain ranges act as obstacles in the path of humid air. This causes an intensification of ascending motion, resulting in the appearance of thick cloudiness and abundant precipitation on the windward side of the mountains. Thus the considerable downpours in Cherrapunji, situated in Bengal at a height of 1300 m above sea level are explained (Figure 46). The high mountain ranges of Central Asia, stretching from west to east, limit the propagation of humid oceanic air masses to the south and southeast, and prevent the penetration of moisture into Soviet Central Asia.

Mountain ranges have a large influence on the distribution of precipitation in North America, where the meridionally lying Rocky Mountains prevent quite effectively the transport of humid air from the Pacific Ocean to the continent. Thus, ocean-facing slopes receive up to 2000–3000 mm of precipitation annually, while to the east of the mountain ranges the annual amount of precipitation is only 300–500 mm or less. This explains why in Vancouver and Sitka, situated to the west of the Rocky Mountains on the coast of the Pacific Ocean (Canada), the amount of precipitation is respectively 1385 and 2161 mm, and on the coast of the Hudson Bay only 350 mm (Port Nelson).

Humid air masses from the Atlantic are carried to Europe without encountering obstacles. As a result, annual precipitation in excess of 500 mm is characteristic both of western and of eastern Europe up to the Ural Mountains.

The influence of mountain ranges on the distribution of precipitation is manifested in many parts of the world. It is clearly manifested in Caucasia. In Batumi, situated on the shore of the Black Sea at the foot of the mountains, 2465 mm of precipitation falls on the average each year, while in the neighboring Kura-Araks lowland and in the Ararat valley only 200–300 mm. This variation in the amounts of precipitation is due to the presence of a number of ranges stopping the moisture on the western windward slopes. Lowlands situated more to the east are supplied with little moisture. Due to the closely lying mountain range on the shore of the Caspian Sea, over 1000 mm of precipitation falls in the region of Lenkoran, whereas on the lowland, situated alongside on the same coast of the Caspian Sea, only about 200 mm of precipitation falls annually.

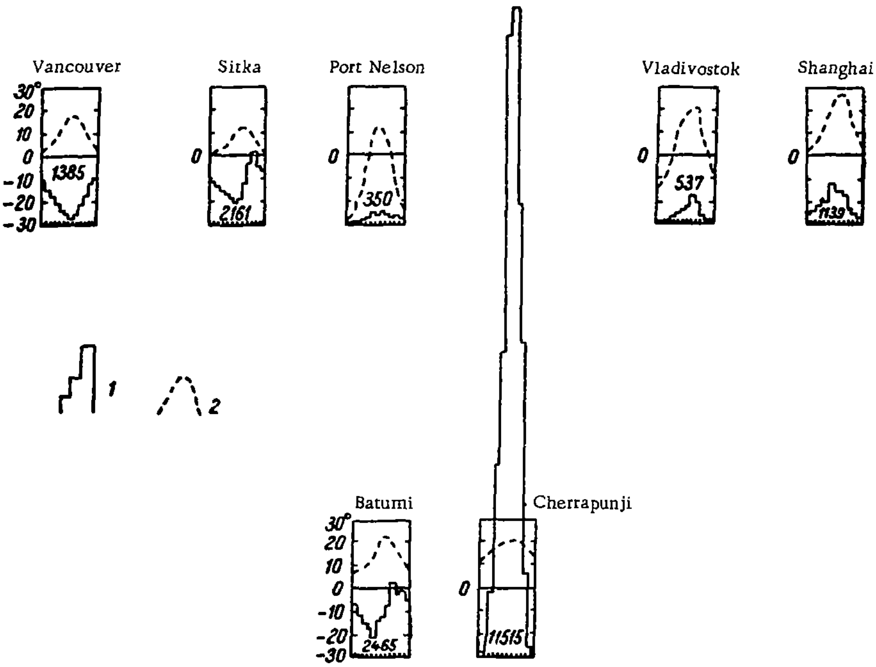

FIGURE 46. The annual variation of the precipitation (1) and temperature (2) under various physico-geographic conditions

To the east of the Caspian Sea the Transcaspian deserts begin, and one may assume that if there was a mountain range 2–3 km high along the eastern coast, the slopes facing the sea would be considerably moistened. In places where there are no mountains, the Caspian Sea exerts only a small influence on the precipitation, since its water surface is too small to moisten appreciably the air flowing over it.

On the surface of seas and lakes there is less precipitation than on the coast. It is known, for example, that even on the islands of the Mediterranean, Adriatic, Aegean, Baltic and a number of other seas, there is more precipitation than on the adjacent water surface. In the Baikal, Sevan, Issyk-Kul and other lakes, pluviometric depressions are clearly displayed which show that on the lakes themselves there is less precipitation than at some distance inland from their shores. Thus, the considerable amount of moisture evaporating from seas and lakes does not go to the formation of precipitation over them. This is due to a number of factors. Chief among them are: (1) the low elevation of sea or lake surfaces as compared with the shores, as a result of which the air flowing from the shores to the water surface descends; this effect is most clearly displayed in the presence of coastal hills; 2) the decrease and often even the stopping of summer thermal convection over seas and lakes, due to the relatively low temperature of the water at that time and the intense heating of the land.

The formation of precipitation is also largely effected by oceanic currents. In the zone of warm currents there is more precipitation. Cold currents inhibit cloud formation, since the temperature inversion, arising as a result of the cooling of the lower air layers, prevents the formation of thick clouds and precipitation. The influence of cold currents on the precipitation regime on the west of South America, the northwest of Africa and other regions, is clearly seen on the precipitation map.

Cyclones and precipitation

The monthly, seasonal, and annual precipitation in middle and high latitudes depends on the type of circulation prevailing: cyclonic or anticyclonic. We recall that overcast and rainy weather is usually related to cyclones, and dry weather – to anticyclones.

When the moisture content of the air and the cyclonic circulation are sufficient, a considerable amount of precipitation often falls. As is shown by the calculations of Z.L. Turketti, a single cyclone may in some individual cases give over 40–50 billion m^3 of water during 2–3 days. This volume of water is sufficient to fill up the basin of Lake Sevan.

It can be seen from the map of the precipitation distribution, that in the North Atlantic and in Europe the largest annual amount of precipitation falls between 65° and 50° NL. The amount of precipitation decreases there from west to east. It can be seen from Figure 47 that in Ivigtut (south Greenland), the annual precipitation reaches 1128 mm, in Reykjavik (Iceland) – 870 mm, in London – 616 mm, in Leningrad – 522 mm, and in Arkhangelsk – 466 mm. Further to the east, the amount of precipitation decreases, and on Lake Baikal (Olkhon Island) the amount is as low as 158 mm annually. To the south of this belt, the annual amount of precipitation decreases in spite of the fact that the moisture content of the air transported over the south is higher than in the belt of maximum precipitation. This precipitation distribution is determined by the circulation features. A study of the frequency of cyclones and anticyclones in the northern hemisphere showed that cyclones are particularly frequent in the northern half of the middle latitudes. Their number decreases from north to south in the belt between 65° and 45° NL. In general, the frequency of cyclones over Eurasia decreases sharply from northwest to southeast. In accordance with this, the largest amount of precipitation falls on the oceans and near the shore, and the smallest amount – in the interior of the continent. Due to the low frequency of cyclones to the south of 50° NL, clear weather prevails there, and consequently the air is heated. Even during cyclones it does not always reach the state of saturation.

The deserts of Soviet Central Asia are a result both of the low cyclonic activity, and of the intense heating of the air and its removal from the state of saturation.

There is no doubt that the distribution of the precipitation over the terrestrial globe would be different if its formation depended only on the moisture content of the air. However, it is related to the character of the atmospheric circulation, and primarily to the cyclonic and anticyclonic activity.

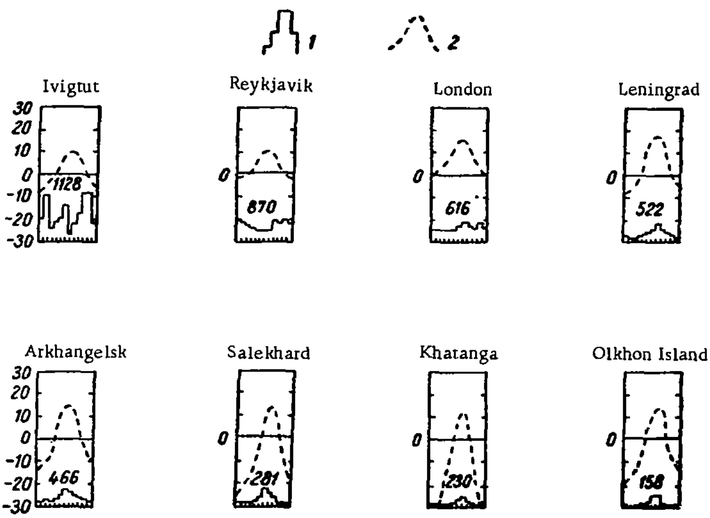

FIGURE 47. The annual variation of the precipitation (1) and temperature (2) in the north of Europe and Asia

At low latitudes, as was said above, precipitation is due not so much to cyclonic activity, as to the convectively unstable stratification of the air. This factor has a particularly strong effect in Southeast Asia, where it determines both the amount of precipitation, and its distribution.

Mountain ranges also cause much precipitation. The appearance of a thick thermal depression over the south of Asia leads to monsoon winds directed from the region of relatively high pressure over the Indian and Pacific oceans to the system of this depression.

Monsoons extend over regions lying to the south of the main mountain ranges of Asia. Because of this, moisture transported from the Indian Ocean precipitates on the south and southeast slopes of these ranges. The transport of moisture from the Indian Ocean does not effect the water cycle in the USSR directly. The narrow belt of high precipitation on the extreme east of Asia results from summer monsoon winds as well as from frequent cyclonic activities developing there. Due to these cyclonic activities moisture is transported from the southern and eastern seas which wash the shores of Asia.

A completely converse situation exists on the other oceanic shore. On the western shore of Africa, where cyclones are very rare, the mean annual precipitation at 25°–30° NL is approximately 100 mm. This is also the case on the Arabian coast of the Indian Ocean and in other places.

It is not true that because of the large amount of evaporation the oceans always receive much precipitation. In a number of oceanic regions where anticyclonic conditions with characteristic temperature inversion prevail (for example, on the Atlantic coast of Africa at 20°–25° NL), the annual precipitation is negligible. Atmospheric drought exists there over the ocean, since the precipitation is not higher than that in the deserts of Soviet Central Asia. The annual evaporation in these regions exceeds 1500 mm, however.

Precipitation and local evaporation

Comparison of the monthly amounts of evaporation and precipitation for a bounded land territory shows that there is no direct relationship between these elements. On the contrary, on the average the relationship is inverse. An increase in precipitation is usually associated with a decrease in evaporation.

Hence it follows that the evaporation does not play a significant role in the formation of precipitation over a bounded land territory. We recall that over such water basins as the Aral and Caspian seas, the precipitation is very small. On the other hand, the annual quantity of water evaporating from the surface of these seas amounts to a layer 800–1000 mm thick. Also air transport from west to east prevails over these seas in the troposphere, due to which the air is moistened even more by the evaporation from the surface of these seas. It would seem that the amount of precipitation at the eastern shores should be larger than at the western shores. In reality the situation is different. This is seen particularly clearly in the case of the Caspian Sea. The eastern shores of the Caspian Sea are considerably less moistened than the western coasts, but the area of the Caspian Sea is over 400,000 km^2, i.e., it is larger than a number of other seas combined. This also indicates that the moisture, which evaporates intensively from relatively large water basins, participates to a small extent in the formation of the precipitation falling both on the water surface and on the nearest land areas.

The determination of the role of the local evaporation in the formation of precipitation is interesting from more than just a theoretical point of view. This problem is of practical importance for large-scale climate improvement works. Therefore when investigating the water cycle in the atmosphere, scientists have recently paid special attention to a quantitative determination of the water cycle elements, and in particular to the determination of the role of evaporation from a bounded land territory in the fall of precipitation on it.

As an example we turn to the beautiful high mountainous Lake Sevan situated in the Armenian SSR. The reader probably knows that for several years now the age-old water reserves of the Sevan have been used to obtain electric energy and to irrigate fields in the Ararat valley and in its northern foothills.

Before water was taken from the lake, its height above sea level was 1916 m, and its area about 1400 km^2. According to calculations of the water balance made by V.K. Davydov, B.D. Zaikov, and improved by co-workers at the Institute of Power Engineering and Hydraulics of the Academy of Sciences of the Armenian SSR, on the average over 1 billion 200 million m^3 of water evaporate annually from this area. Twenty-eight rivers flowing into the lake bring about 0.77 billion m^3 of water each year. The annual amount of precipitation on the surface of the lake is over 0.55 billion m^3. Consequently, the annual inflow of water to the lake is 1.320 billion m^3. Only one river flows from the lake — the Razdan.

Before water was taken from the lake, it lost about 0.05 billion m^3 of water annually because of this river. If one takes into account that up to 0.09 billion m^3 of water escape annually by filtration from the lake, one finds that over 1.2 billion m^3 evaporates annually from the surface of the lake, i.e., 91% of the water flowing into the lake evaporates.

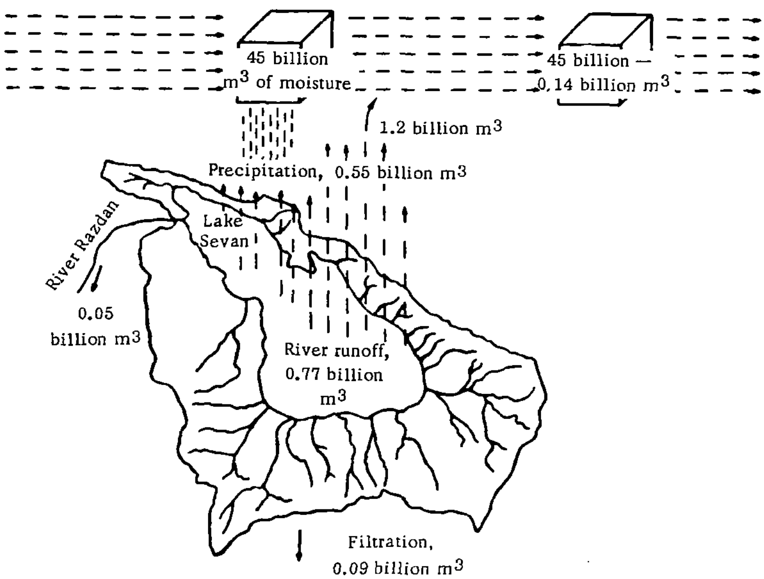

FIGURE 48. Components of the water cycle in the basin of Lake Sevan

What portion of the evaporating water falls on the basin of the lake and on its surface? Calculations performed during 1947 by G.D. Zubyan and by the author showed that the air masses flowing over the lake transport approximately 45 billion m^3 of water annually. About 1.2 billion m^3 evaporates. Thus the water evaporating from the surface of the lake amounts to only slightly over 2.5% of the whole moisture carried by the air. If one takes into account that Lake Sevan is situated in a valley surrounded by mountains, it becomes obvious that some part of the evaporating water falls as precipitation on their slopes. However, this amount of precipitation, according to calculations, is only a small fraction of the precipitation formed from the water vapor carried by the air from the outside.

Consequently, as shown in Figure 48, of the 45 billion m^3 of moisture transported by the air over the lake, only 0.14 billion m^3 are retained there, leaving the lake via the river (0.05 billion m^3) and by filtration (0.09 billion m^3), while a considerable portion of the precipitation evaporates from the lake and together with the atmospheric moisture is carried beyond the boundaries of Transcaucasia.

The amount of precipitation falling on a bounded land territory is almost independent of the amount of moisture evaporating from this territory. This can be seen from the graph (Figure 49) of the annual variation of the evaporation from the surface of Lake Sevan and of the annual variation of the precipitation on the Island Sevan, situated in the western part of the lake.

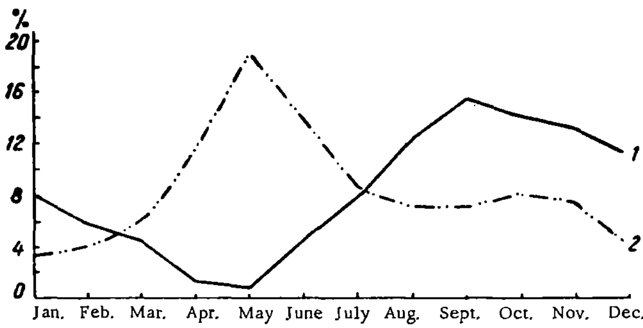

FIGURE 49. The annual variation of the evaporation from Lake Sevan (1) and the annual variation of the precipitation on the Island Sevan (2)

It can be seen from this figure that the variation of the precipitation curve is almost the inverse of the variation of the evaporation curve. The lowest evaporation is observed during the spring (April–May), the highest evaporation – during the fall (September–October). On the other hand, the largest amount of precipitation in the lake basin falls during the spring.

The main idea behind removing a large portion of the water from the lake is to reduce the evaporation considerably, and thus escaping moisture, and use it for irrigation and for generating hydroelectric power.

However, since there are now new ways of producing energy (thermal power stations using natural gas, interconnection of the power systems of the Transcaucasian republics, etc.), we can now reduce the water discharge from lakes for power generation and from 1965 use this water for irrigation (which requires less water) thus preserving Lake Sevan.

According to the new scheme, the level of the lake will drop by an additional 7–8 m until 1965, bringing its total drop to 20 m, i.e., the level of the lake will be 1896 m above sea level instead of the 1916 m before the beginning of the project. Due to some decrease in the surface area of Lake Sevan, and consequently, also in the evaporation, the runoff amounts to 171 billion m^3 each year.

In order to reduce the evaporation, the Institute of Water Economy of the Academy of Sciences of the Armenian SSR carries out experiments to cover the water surface by a special film. Independently of the successful attempts to reduce evaporation and to produce artificial rain, the rerunoff from the Sevan will amount to 171 billion m^3 after 1965; this is sufficient for irrigating the lower lying land.

For the operation of the hydroelectric stations of the Sevan – Razdan [Zenga River] cascade the transfer of 300 billion m^3 of water each year over an underground runoff from neighboring bases will be investigated. This project has already begun.

We now discuss the new methods for calculating the water cycle.

Calculations of the water cycle

During 1949–1950 M.I. Budyko, O.A. Drozdov, K.I. Kashin, and Kh.P. Pogosyan proposed a new method for calculating the water cycle. Use was made of modern data of aeronautical observations. In order to calculate the water cycle, the role of the most important factors taking place in this complex process must be determined. These factors are the transport of moisture in the atmosphere, precipitation, runoff, and evaporation.

Clouds giving precipitation at middle latitudes lie mostly at heights of 5–6 km. Since the moisture content per unit volume of air rapidly decreases with height, at the 5-km level the amount of moisture is so small that it practically plays no significant role in the formation of the precipitation. Therefore, in calculations of the amount of moisture transported over any territory, one takes into account only the amount of moisture contained in an air column 4–5 km high.

As the air gets farther from the shores of the ocean, the amount of water vapor contained in it decreases gradually due to the fall of precipitation. The more the precipitation that falls along the path of the air, the faster the decrease in the moisture content. If all the falling precipitation would leave through the river runoff, then, as we saw, the air would rapidly become dry. In reality this does not occur since a considerable part of the fallen precipitation returns to the atmosphere through evaporation. However, the role of evaporation in the fall of precipitation on territories of different sizes is different. Since the air is in continuous motion, then for the same wind velocity the moisture evaporating from a small area will be carried beyond its boundaries faster than the moisture evaporating from a large area. Therefore the moisture evaporating from a small area will in most cases not have time to cause an increase in the precipitation on the given territory.

Even moisture evaporating from a large area can increase the amount of falling precipitation to some extent if the atmospheric circulation causes the upward motion of this moisture; this usually occurs in the case of cyclonic activity. In anticyclonic circulation, on the other hand, when the evaporation is most intense, the evaporating moisture cannot as a rule play an important role in increasing the amount of falling precipitation.

As an example of an approximate calculation of the water cycle over a certain region we give the calculation of the water cycle for the basin of the River Oka, performed by the author jointly with K.I. Kashin.

The area of the Oka basin is about 240,000 km^2; along the parallel the basin stretches approximately 600 km, and along the meridian – 400 km. The mean annual amount of precipitation in the Oka basin is approximately 550 mm, i.e., 134.2 km^3 of water. The runoff of the Oka at the gage line of Novinki amounts on the average to 33.2 km^3 annually. Consequently, the annual evaporation from the entire basin is 101 km^3.

The calculation shows that the amount of moisture evaporating from this territory is considerably smaller than the annual amount of moisture which is transported by the air over this territory. Thus, for example, the annual amount of water carried by the air over the basin of the River Oka is about 1285.0 km^3. On the basis of the above given data on the precipitation (134.2 km^3), runoff (33.2 km^3), and evaporation (101.0 km^3) we see that as compared with the annual amount of water carried by the air, the precipitation amounts to approximately 10.5%, the runoff – to 2.6%, and the evaporation – to 7.9%.

Hence, even for relatively large basins, the precipitation, runoff, and evaporation amount to a small fraction of the water which is transported by the air during the year. We thus conclude that the evaporation from the surface can play an important role in moistening the moving air only when it takes place over very large areas. According to the above results, the evaporation amounts to 7.9% of the moisture transported by air while the air mass moves a distance equal to the length of the basin (600 km). This is the amount of water which returns to the air after the fall of the precipitation. It is natural that the larger the distance traversed by the air mass, the larger the role of the evaporation.

The same authors made a calculation of the water cycle for the European territory of the USSR. According to these calculations, 8507 km^3 of water are transported annually by the air over the European territory of the USSR. The area of the basins of all the rivers of the European territory of the USSR is about 6.5 million km^2. The annual runoff on this territory is 928 km^3 of water, and the mean annual amount of precipitation can be taken as 480 mm, or 3120 km^3. Consequently, in this case 2192 km^3 of water evaporate. It follows from the values obtained for the runoff, precipitation, and evaporation that, with respect to the annual amount of moisture transported by the air, the precipitation amounts to about 37%, the runoff – to 11%, and the evaporation – to 26%.

These data indicate the role of each component of the moisture balance on such a large area as the European territory of the USSR. As can be seen, from the immense amount of moisture (8507 km^3) transported mainly to the west, the river runoff accounts for only 928 km^3, i.e., about 11%, and the greatest part of the moisture (7579 km^3) is carried beyond the boundaries of this territory, mainly in the eastern direction and irrigates the lands lying behind the Urals.

Calculations give the following results on the role of the evaporation. Of the entire moisture evaporating from the European territory of the USSR, only 13% falls again on the same territory in the form of precipitation. The remaining 87% of the precipitation is formed from moisture brought there from the outside.

These conclusions are correct for mean annual conditions. New investigations, conducted by M.I. Budyko, O.A. Drozdov, A.I. Burtsev and others, showed that in individual short-time intervals the role of the local evaporation in the fall of precipitation can be considerable. This is possible when little moisture is carried beyond the boundaries of the territory under consideration and most of the evaporation moisture goes in the formation of clouds and precipitation. Such conditions are sometimes created during the summer half year, but not frequently.

Returning to the problem of the mean annual distribution of the precipitation over the terrestrial globe, it should be noted that the mean monthly and annual precipitation amounts conceal very important details. Whereas the mean temperature for an individual month or year does not differ very much from the many-yearly means, i.e., from the norm, this is not true for the precipitation. The precipitation amounts during individual months or years all over the globe vary widely — from zero to twice or three times the norms and more. Often, even the daily amount of precipitation exceeds the monthly amount and in individual cases also the seasonal norm. Cases have been recorded when in tropical countries the amount of precipitation that had fallen during 24 hours exceeded 1000 mm, i.e., a water column 1 m high. In Moscow, for example, the maximum annual amount of precipitation ever recorded was 834 mm, the minimum — 272 mm, with a mean annual norm of 617 mm. (Interesting information on the climate of Moscow may be found in the booklet: Kolobov, N.V. Klimat Moskvy (The Climate of Moscow). — Izdatel'stvo "Moskovskii rabochii." 1960.)

In the USSR the largest daily amount of torrential precipitation in the southern regions reaches 200 and even 300 mm, whereas in the northern latitudes it does not exceed 70–80 mm.

Sometimes even in the steppe and forest-steppe zones no precipitation falls during a month; this leads to an atmospheric drought, so harmful to nonirrigated farming. In the subtropical deserts there are regions where for some successive years there is no precipitation. Such a region is the Peruvian coast of South America.*

* An interesting description of the climate of these regions can be found in the book by the well-known Czechoslovakian travellers: Hanzelka, I. and M. Zikmund. Cherez Kordil'ery (Through the Cordilleras). —Izdatel'stvo "Molodaya gvardiya." 1960.

ATMOSPHERIC FRONTS

Variation of the physical properties of moving air masses

The physical properties of the air in the troposphere are not uniform. This nonuniformity, as we have seen, is due not only to the nonuniform distribution of the solar energy over the terrestrial globe, but also to the influence of the underlying surface, which possesses different properties. Air masses, moving over various conditions of the underlying surface, change their physical properties; they acquire new ones, which are characteristic of the geographic regions over which they move. The variations in the physical properties, or, the transformations of the air, take place continuously, since the air is rarely in a state of rest.

When air undergoes transformations its temperature and humidity vary.

It should be mentioned that the process of air transformation is one of the most difficult problems in meteorology; nevertheless some successes have been achieved recently.

Air moving from north to south is usually heated, thus being further removed from the state of saturation. However, at the same time the air acquires the capacity to absorb more moisture from the underlying surface and thereby increase its moisture content. When moving from south to north, on the other hand, the air is cooled, and consequently becomes saturated even when its moisture reserve is low.

Under appropriate circulation conditions, part of the moisture may fall in the form of precipitation. In this case the moisture content of the air decreases. Under the same circulation conditions the moisture content of the air increases more rapidly when it moves over water surfaces (oceans, seas, etc.) and more slowly over a low-humidity underlying surface.

In addition to continuous variations of the temperature and humidity of the air caused by its constant heat and moisture exchange with the underlying surface, the optical properties of the air also vary. The latter is caused by the presence of various impurities in the form of very small particles of dust and of combustion products suspended in the air. The air becomes less transparent when it contains a large amount of impurities. The turbidity of the air is usually increased over deserts and industrial regions.

Investigations have shown that air masses, being in continuous motion, modify their properties continuously.

According to their temperature, air masses are divided into two classes: warm and cold. "Warm" is the term used for an air mass arriving on a relatively cold underlying surface, while "cold" – for an air mass arriving on a relatively warm underlying surface. It is natural that cold air masses are formed at high latitudes, and warm masses – at low latitudes.

Relatively uniform masses of air, extending several thousands of kilometers horizontally and several kilometers vertically, are called air masses.

The formation of air masses takes place when they become stationary over some regions, which is therefore called an air-mass formation focus.

It was assumed that an air mass forms and moves as one unit. A classification of the air masses according to a geographic criterion was therefore proposed. According to the latitude zones of the northern hemisphere one distinguishes arctic, polar, tropical, and equatorial air. Polar air is the air of temperate latitudes; the remaining three types correspond to the names of the latitudinal zones.

Since there are continents and oceans in each latitude zone (except the equatorial one), an additional criterion was used to divide air masses into marine and continental; the differences between them are in the first place in the different moisture content.

In the 1930's the geographic classification of the air masses was a matter of interest to many forecasters and meteorologists. In accordance with this classification, mean meteorological characteristics were calculated for various air masses at various points of the Earth, thus facilitating the discrimination of air masses. Another important aspect was the determination of the boundaries between air masses; as we shall see below, the greatest weather variations occur where different air masses meet.

It was later established that air masses do not have conservative (almost nonvarying) properties, since, being in constant motion, their physical properties vary continuously.

It is obvious that the air masses are not being formed at once, as was assumed earlier, but undergo continuous transformation, since an air mass can form as a whole at focuses in the absence of motion. On the other hand, an air mass of small vertical extension can in rare cases lie still over one territory.

During the period when no data of systematic aerological observations were available, the geographic classification of air masses played a positive role, being the only accessible data.

Meteorologists now have at their disposal extensive results of direct observations at the surface of the Earth and at heights, carried out at a wide network of stations. The mean characteristics of air masses are therefore no longer used. With the new data of aerological observations, the actual physical properties of air masses can be determined and in each individual case the variations in these properties followed during the motion.

It is not enough to say that the air is in constant motion, since by this it may be understood that the whole mass, from the surface of the Earth to large heights, moves as a single whole in one direction. In reality the air motion is very complicated, since the wind direction and velocity vary considerably with height. As a result, air particles move in different directions at different levels. If we take into account that in addition to the horizontal motion, there is vertical motion of air masses and turbulent motion of air particles, we get some idea of the chaotic motion in the atmosphere.

The reader has probably observed more than once how, in a calm or in a weak surface wind, the clouds move with an appreciable velocity in the opposite direction to that of the wind at the surface of the Earth. It also often occurs that clouds situated at different levels move in different directions. This complex system of motion of the air masses considerably complicates the study of the variation of the physical properties of air.

With all the chaotic character of the motion in the atmosphere, air masses moving from some regions into others, tend to acquire the properties characteristic of that air which stays for a long time at the given latitude and over the given underlying surface.

In addition to observing the variation in the properties of the moving air masses, it is also important to estimate quantitatively the transformation of air moving over different underlying surfaces. First attempts have already been made to calculate the temperature distribution in the lower atmosphere on the basis of the astronomical characteristics of the Earth and on data of the optical properties of the underlying surface.

Theoretical computations of the temperature distribution at heights closely agreed with the observed one. But these are only the first steps. Many difficulties must still be overcome before one can predict the magnitude of the variations of various properties of air moving over extensive spaces of continents and oceans at the surface of the Earth and at heights. Many failures in weather forecasting even 1–3 days ahead result from the present inability to calculate accurately and in time the magnitude of one of the principle transformation elements of air masses — the temperature variation. This is due to the fact that the thermal transformation of the air affects more than the direct variation of the weather elements. Thermal transformation, extending over considerable layers of the troposphere, exerts a direct influence also on the structure of the pressure field, and consequently, on air currents and on the life of cyclones and anticyclones.

The longer the air masses stay over a relatively uniform underlying surface, the more they acquire new properties characteristic of the given region. The rate of the air-temperature variation depends on the difference between the temperatures of the air and of the underlying surface. The larger this difference, the faster the air is heated or cooled. By turbulent mixing the received heat or cold is transferred to higher-lying layers of the troposphere. However, when the surface air layer is heated and instability develops, the heat transfer to the upper layers is faster than when the surface layer is cooled. In the latter case, the arising temperature inversion does not contribute to intensive turbulent heat exchange with the higher-lying air layers.

The variation of the air temperature with height is not only caused by heat transfer from the underlying surface, but also by vertical motion, giving rise to adiabatic cooling and heating.

It was determined that the temperature of moving air varies considerably. Thus, for example an air mass having traversed the distance from the region of Tikhaya Bay to the south of the European territory of the USSR in four spring days, was heated at ground level by 16°C, at the height of 3 km by 6°C, and at a height of 5 km by only 4°C. As a result its stratification changed (Figure 50). In the region of Tikhaya Bay the air had a stable stratification; at ground level temperature inversion was observed in the layer up to 1.5 km, i.e., isothermal conditions, and the mean lapse rate did not exceed 3.0°C per km. To the south of the European territory of the USSR the lapse rate was already 5.3°C per 1 km, and slight instability was observed in the lower air layer. In an opposite case, during the three days of motion of an air mass in spring from the region of Lvov to the region of Naryan-Mar, its temperature at ground level fell by 11°C. In the first 500-meter layer inversion appeared. From a height of 2 km the cooling of the air mass was decreased as a result of the inversion and the decrease in the turbulent mixing. In fact, at a height of 3 km the temperature fell by 2°, and at the height of 5 km it remained almost unchanged.

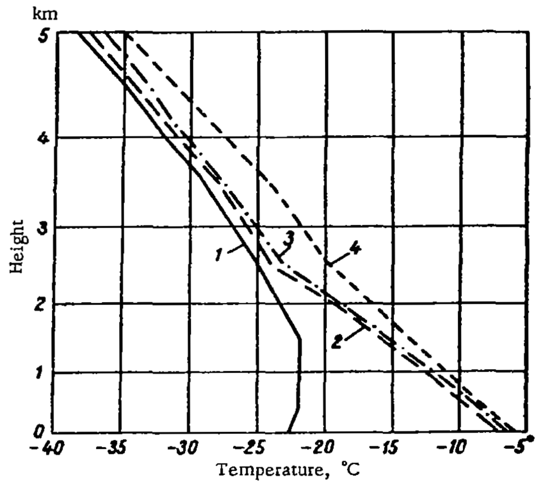

FIGURE 50. The temperature variation of air moving from Franz Josef Land to the European territory of the USSR from 29 March to 2 April, 1939

1 – Tikhaya Bay, 29 March; 2 – Slutsk — Murmansk, 31 March; 3 – Minsk — Smolensk, 1 April; 4 – Kharkov — Rostov, 2 April.

Thus, whereas in the first case an air mass which was initially stable became unstable owing to the large heating of its lower layers, in the second case it became stable owing to the cooling of the lower layers. Since the variation in the stratification of an air mass affects the weather, a series of physically explicable qualitative propositions have been made, which are used by forecasters in their work.

1. When air is transported from a cold water surface to a heated continent, layer clouds or fog observed over the sea are scattered on the heated land. As the air is heated and instability develops, convection clouds (cumulus clouds) may appear in the lower layers. Conversely, in motion from the hot land to the cold water surface — the instability decreases, cumulus clouds are dispersed and fog or low-layered clouds may form.

2. When stable stratified air is transported from a cold land to a warm water surface, layer clouds or fog observed over the land disappear over the sea. Conversely, in the case of motion from the sea to the land, fog or low-layer clouds are formed.

3. When vertically unstable stratified air is transported from a cold land to a very warm water surface, it gets heated over the sea, its instability increases, cumulus clouds appear which pass into cumulonimbus clouds, and torrential rains fall. In the case of a reverse air motion, the convection phenomena are weaker and the clouds disappear.

4. If air is transported from a warm water surface to a very hot land, then clouds that were weak convection clouds over the sea develop while moving over the continent and torrential rains fall. In the case of motion from land to a water surface, on the other hand, thick convective clouds gradually disappear.

In all these cases the air must contain a sufficient amount of moisture, since, when it is very dry, its heating and cooling do not give rise to sharp weather changes. If the air encounters obstacles over the land the above processes develop more strongly. This is most often observed on sea shores, in particular on the Black Sea shore of the Caucasus.

The humidity variation of moving air masses is also of considerable importance. When the temperature rises, the moisture content of the air may rise if the motion takes place over an extensive water surface or humid soil. During the winter, for example, when air masses move from the cold continent to the warm ocean, the air temperature and its moisture content rise simultaneously. Warm air masses absorb the moisture evaporating from the ocean like a sponge; conversely, when moving from the ocean to the cold continent, air masses cool down, become saturated, and clouds and precipitation are formed. It is natural that in these cases the air loses a part of its moisture.

In anticyclonic activity, particularly when there is extra heating, the air masses are enriched with moisture by evaporation. In the absence of evaporation sources, as, for example, during summer in waterless steppes and deserts, the air, heating up, gets further from the saturation point. In these cases, even when the evaporation is strong and the moisture content is increased, the relative humidity at daytime hours drops to 20–10% and lower.

Thus, the variation in the moisture content of the air takes place, on one hand as a result of the almost continuous evaporation from the underlying surface, and on the other hand, as a result of cloud formation and fall of precipitation. The quantitative allowance for the variation in the moisture content in air masses is very important for the correct prediction of precipitation. The greatest failures in weather forecasting are related to precipitation forecasting.

Other phenomena, for example the turbidity of the air, are not as important as the temperature and humidity.

Sharp temperature variations at middle latitudes are observed during all the seasons. Rapid interchanges of cold and warm weather have often been observed. During the winter half year cold air masses penetrate to the European territory of the USSR, usually from the north or northeast, the coldest air arriving from the east, from Siberia. The particularly severe frosts in eastern as well as in western Europe, are connected with the penetration of cold air from Siberia. Thus, for example, even in France the temperature during the winter of 1929 fell, upon the penetration of air from the east, to -10°, -15°C.

The map of the monthly mean temperature deviations from the norm during February 1929 shows that in western and Central Europe the monthly mean air temperature was lower than the norm by 5°–10°C (Figure 51). The frosts of January 1940 were also connected with the penetration of cold air from the north of Siberia. In Moscow the air temperature dropped to -42°C, and in Klin – to -52°C.

In the case of a prolonged displacement of air masses from the west, the temperature rises and even in the middle of the winter in the central regions of the European territory of the USSR, thawing begins. Warm air masses arrive not only from the direction of the Atlantic in the west, but also from the far south. During the winter in western and eastern Siberia, they usually penetrate from northwest, west, and southwest. However, due to their long stay over the cold continent, the air masses lose a part of their heat. Therefore in the middle of the winter the temperature does not reach zero and thawing does not begin. Spring cold waves , as well as early fall frosts, are usually observed upon the penetration of air masses from the north, almost along the meridians. During the summer, air masses with a high temperature and a low moisture content arrive from the heated continent in the east, and humid air with a moderate temperature – from the direction of the Atlantic in the west.

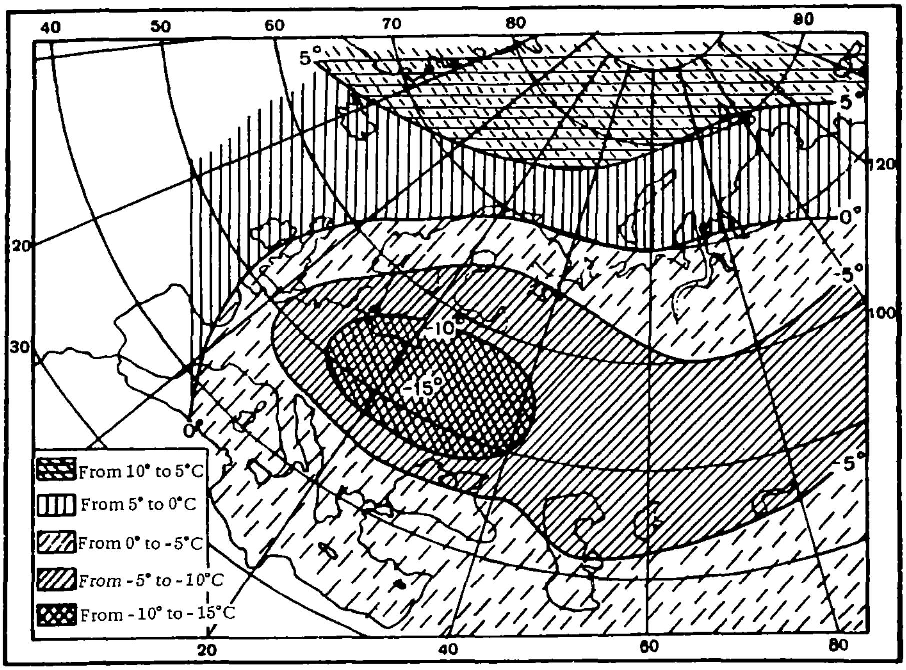

FIGURE 51. Deviation of the monthly mean air temperature from the norm in Europe, February 1929

Therefore, when western transport prevails in the northern and central parts of the European territory of the USSR, the summer is cool and rainy.

Frontal zones in the troposphere. Atmospheric fronts

We have seen that the nonuniform heating of the surface of the Earth and of the air in the troposphere is the reason for the appearance of horizontal temperature and pressure gradients and for the formation of air currents.

Due to the transport, air masses of different properties may get closer or farther from one another. When air masses with different physical properties approach one another, the horizontal gradients of the temperature, humidity, pressure, and other meteorological elements increase, thus resulting in an increase in the wind velocity; conversely, as they get farther from one another, the gradients decrease. Those zones in which different air masses approach (for example: relatively dry and cold, on one hand, and humid warm on the other hand), are called transitional or frontal zones. In frontal zones there is a struggle between the cold and warm air masses. As a result of this struggle the cold air masses are broken through in the regions where the warm masses are situated, and the warm masses penetrate to the regions where the cold masses are situated. Thus various air masses gradually acquire properties characteristic of the air of the given region.

Tropospheric frontal zones can be observed every day in the temperature and pressure field mainly at extratropical latitudes where the inflow of solar energy to the north and to the south of the temperate zone is different. The horizontal temperature and pressure gradients are larger there than anywhere on the terrestrial globe. Frontal zones continuously appear, become sharper, and then break down. However, they may have different intensities, depending on the temperature difference between the approaching air masses.

When frontal zones intersect in the lower layers of the atmosphere the temperature, pressure, and humidity rapidly fall from the warm to the cold air depending on the large horizontal gradients, and high wind velocities are observed. At middle latitudes at heights of 10–12 km in these zones, the winds often reach hurricane force, i.e., 200 km/hr and more. As we see below, frontal zones play an important role in the development of atmospheric processes.

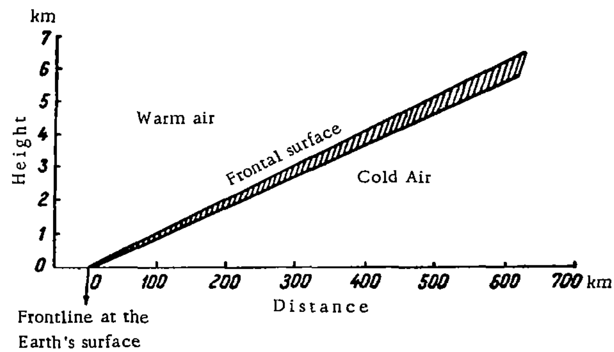

FIGURE 52. Diagram of the interface between cold and warm air masses

Since cold and warm air masses have different densities, they lie on each other with some inclination to the vertical. The cold air, being denser and heavier, is wedged in under the warm, lighter air. In this boundary zone between air masses of different properties, cyclones and anticyclones often arise, bringing both bad and good weather.

The dimensions of the transitional zone are small as compared to those of the air masses. Interfaces appear in a frontal zone between cold and warm air masses; these are called atmospheric fronts. Frontal surfaces are always inclined toward the cold air, which lies under the warm air in the form of a narrow wedge (Figure 52). The inclination angle of the frontal surface to the horizon is very small, smaller than 1°, and the tangent of the angle varies between 0.01–0.02. This means that if one moves a distance of 200 km away from the frontline at the surface of the Earth in the direction of the cold air, the frontal surface will be at a height of 1–2 km. Moving in the horizontal direction a distance of 500 km, the frontal surface will be at a height of 2.5–5.0 km. Since the inclination angles of the fronts are very small, to represent the fronts more clearly in the vertical plane, the horizontal scale is usually taken many times smaller than the vertical one. In Figure 52 the vertical scale was increased by a factor of almost 50.

The maximum vertical extension of fronts at middle latitudes is 8–12 km. They often reach the tropopause. According to investigations by E. Pal'men, G.D. Zubyan and others, fronts are also observed in the lower layers of the stratosphere.

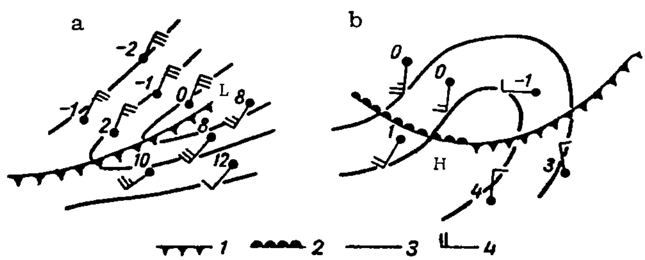

FIGURE 53. Front sharpening (a) and occulusion (b) conditions at ground level

1 – cold front; 2 – warm front; 3 – isobars; 4 – wind velocity and direction.

On tropospheric fronts multilevel cloudiness usually develops, from which precipitation falls. Fronts are most sharply displayed in cyclones, where ascending air motion prevails. The frontal cloudiness of anticyclones scatters due to the descending motion.

Atmospheric fronts are divided into cold and warm fronts.

Cold front is the term applied to a front moving toward high temperatures. After the passage of a cold front the air cools. Warm front is the term applied to a front moving toward low temperatures. After the passage of a warm front the air warms up.

In the temperature and wind field, fronts are most sharply displayed at the surface of the Earth in the system of developing cyclones and barometric troughs. The convergence of the air currents in the frontal zone at the surface of the Earth contributes to this since, as a result of this convergence, air masses with low and high temperatures meet in the frontal zone. Figure 53 a shows the pressure, wind, and temperature field in the trough of a cyclone at the surface of the Earth. The front intensifies, since to the north of it there is a cold air mass with temperatures of 1°–2°C below zero, and to the south of it – a warm air mass with temperatures of up to 10°–12°C above zero.

Fronts at the surface of the Earth become less intense in anticyclones, since the system of air currents diverges (Figure 53 b). In the first part of the ridge the cold section of the front at the surface of the Earth diffuses since the currents point not to but from the front. In a developing cyclone the air tends to move upward and as a result of the dynamic cooling and condensation clouds appear and precipitation falls. In a developing anticyclone, the air is in descending motion and as a result of the dynamic heating the air gets farther from the state of saturation, clouds scatter, and precipitation stops.

The velocity of a front depends on the velocity of the normal component of the wind, which varies within wide limits. In Europe, in the transitional seasons of the year, the mean velocity of fronts reaches approximately 30 km/hr, which is about 700 km per 24 hours; but in a system of cyclones fronts often travel a distance of over 1200–1500 km in a 24-hr period. In these cases, a front which is situated for example in west Europe, is already in the central regions of the European territory of the USSR after 24 hours. If the air currents are parallel to the front, the front hardly moves. Since the temperature and pressure gradients during winter are considerably larger than those during the summer, the winter activity of fronts is distinguished by a high intensity.

It has already been said that in the zone of an atmospheric front, particularly in a developing cyclone, the air ascends and is adiabatically cooled, clouds and precipitation being formed. The air rise takes place not only in the surface layer, but also at heights. Whereas in the surface layer it is caused by the convergence of the surface wind, at heights it is caused by the unsteady motion and the difference in the velocities of the air behind the front and ahead of it.

In the case of a cold front, the rapidly moving cold air behind the front, flowing under the warm air, forces it to move up. As a result, if the dynamic conditions give rise to a general ascent of the air, the warm air begins to slide upward along the inclined surface of the front and is adiabatically cooled.

In the case of a warm front under the same conditions, the warm air also ascends over the wedge of the cold air. The larger the temperature difference between the cold and warm air, i.e., the sharper the front not only at the surface of the Earth but also at heights, the more intense the ascending motion of the warm air, the condensation, and the formation of clouds and precipitation.

On a well-pronounced front there are usually clouds of all levels. Clouds of a warm front may be very thick, and very often extend horizontally to 500–700 km ahead of it, and vertically – up to 6–8 km and more. The length of such a front may reach 1000–2000 km. Even during the summer the upper part of thick frontal clouds is situated in the zone of negative temperatures, and usually consists of ice crystals. Figure 54 shows a vertical perpendicular cross section of the system of clouds characteristic of a warm front. These clouds belong to the layered forms and are situated mainly in the warm air over the frontal surface. The highest clouds (fleecy and fleecy-strati clouds) are situated at heights of 6–8 km. They are the forerunners of a warm front. The appearance of these clouds several hours before the approach of the precipitation zone indicates weather worsening.

Fleecy-stratus clouds are replaced by high-stratus clouds, through which the Sun can still shine; in spite of this they have a large vertical thickness. Further on come denser stratus-rain clouds, giving continuous precipitation which reaches the ground. Below all these lie the stratus and fracto-rain clouds, the height of whose lower boundary may vary from zero to several hundreds of meters, depending on the moisture content. In this case, as can be seen in Figure 54, the low-level clouds are formed not only in the warm air over the front but partially also in the cold air in the immediate vicinity of the frontal surface. The arrows in this figure indicate the direction of the air currents in the warm and cold air under the general transport from left to right of the plane of the diagram.

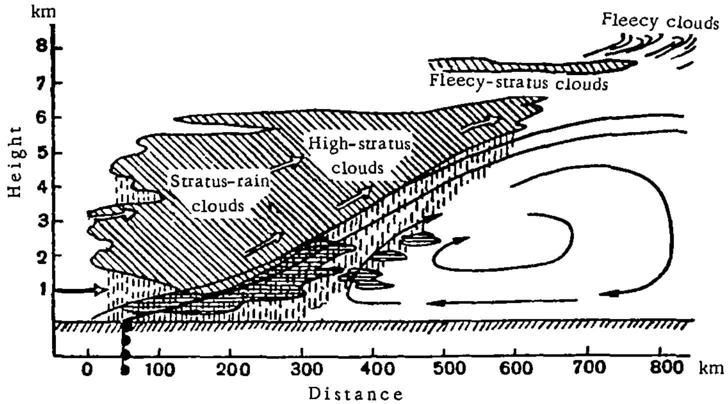

FIGURE 54. Cloud system of a warm front

The cloud system of a thick cold front is shown in Figure 55. As is easily noticed, the profile of warm (Figure 54) and cold (Figure 55) fronts differ appreciably from one another. This is because due to the friction at the terrestrial surface, the lower layer of moving warm air expands in a direction opposite to the motion. A cold front, on the other hand, becomes steeper due to the friction in the lower 1–2-kilometer layer.

The cloud systems of warm and cold fronts shown in Figures 54 and 55 represent situations where the vertical extension and temperature contrasts of the front are large, and intensive ascending air motion takes place. Air masses on both sides of the front are stable. If under all these conditions the cold air is unstably stratified, then the cold front is followed not by stratus-cumulus clouds, but by thick cumulus and cumulonimbus clouds. If at the same time both the cold air and the warm air are unstably stratified, then a thick squall of cloudiness is formed (Figure 56) giving strong torrential rains associated with thunderstorms and sometimes with the fall of hail.

The cloud system of a warm front may also have various forms. In the case of instability of the warm air, convective clouds are formed and torrential rains fall. We assume that the moisture content of the air is sufficiently high.

However, the vertical extension of atmospheric fronts is not always large, and often does not exceed 1–3 km. Accordingly, also the frontal cloudiness acquires a limited extension, with the exception of cases when, owing to instability, convective clouds reaching a height of 5–6 km are formed. Even when the vertical extension of the front is large, the frontal cloudiness does not constitute a continuous medium (see Figures 54 and 55),

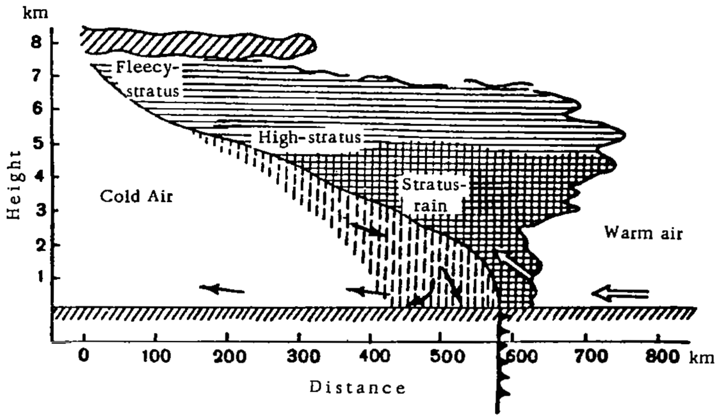

FIGURE 55. Cloud system of a cold front in the case of stable stratification of the air masses

but consists of a series of layers with cloudless spaces between them (Figure 57a). This is due to the fact that in many cases the general ascent of the warm air is upset and that there are alternate layers with ascending and descending air motion in the frontal zone. The latter cause the collapse of the cloud system of the front, up to a complete scattering of the clouds. When the air is very dry, either no clouds are formed at all on the front, or thin clouds of middle and upper levels appear, which do not give precipitation (Figure 57b).

In addition, there are other forms of fronts, which arise as a result of an encounter between a cold and warm front. Joining of fronts occurs as a result of their moving with different velocities. In a cyclonic system cold fronts move as a rule with higher velocities than warm fronts. Therefore when a cold front overtakes a warm front, it joins it, forming a unified front, or, as it is usually called an *occlusion front*. At the beginning the cloud systems of both fronts are preserved and give abundant and continuous precipitation. However, the intensity of the occlusion front gradually weakens as a result of the already operating process of its erosion. The thick cloud systems begin to scatter and the front is observed only in the field of the surface wind by the cloud remainders. Figure 58 shows schematically the encounter of cold and warm fronts moving from left to right. The denser cold air is wedged in under the warm air.

When meeting mountains, all the types of fronts leave much moisture on the windward side. However, as they pass a high obstacle the cloud system of fronts is destroyed, and on the lee side of the mountains the clouds spread out and precipitation often stops. Only after having passed the obstacle, the cloud system of fronts is again restored.

The study of atmospheric fronts must be continued because of its importance practically, particularly for aviation purposes, since thick clouds, as well as sharp weather variations, are connected with fronts. Their study is therefore one of the most important problems in meteorology.

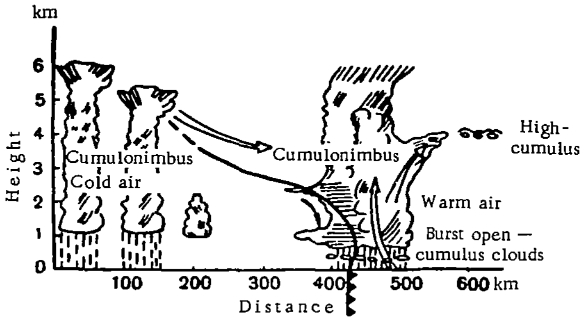

FIGURE 56. Cloud system of a cold front in the case of unstable stratification of the air masses

In spite of their importance, the knowledge of the conditions for front formation is far from being sufficient. This refers first of all to the formation and evolution of frontal cloudiness. The above diagrams only give a general idea of frontal clouds. In reality, clouds in the zone of atmospheric fronts constitute both a continuous medium, and thick layers with cloudless spaces between them.

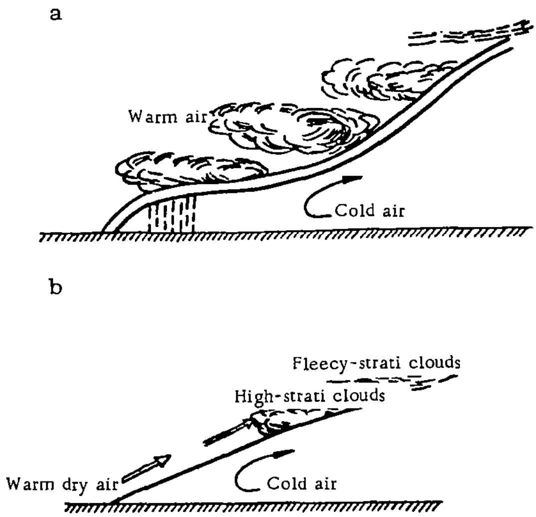

FIGURE 57. Cloudiness stratification of a warm front upon its occlusion (a); cloudiness of a warm front in the case of very dry air (b)

The difficulties in studying the physics of cloud formation on fronts are due to the lack of methods for a detailed study of the peculiarities of clouds under some definite synoptic conditions. This is because a prolonged stay at heights is necessary, which is technically difficult to accomplish.

Modern airplanes flying at high velocities make is possible to carry out observations and various measurements along the airplane's path. Balloons are the most convenient way of studying clouds, but they cannot always penetrate into the cloud in question. In particular, a balloon cannot penetrate into a thunderstorm cloud, since it may catch fire from the lightening flashes.

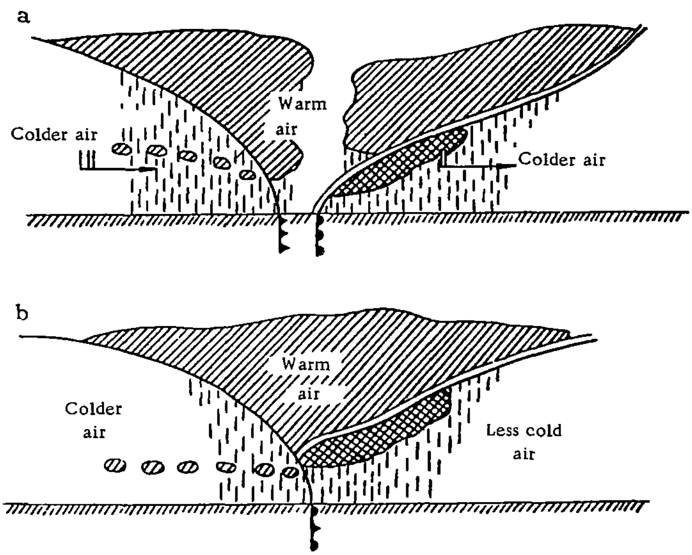

FIGURE 58. Encounter between cold and warm fronts

a – before the encounter; b – after the encounter.

It has already been said that the formation of clouds is caused by the condensation of water vapor resulting from the ascent of the air and its adiabatic cooling. To give an idea of the difficulty of studying the evolution of clouds, it is sufficient to say that the vertical motion of air, causing the formation and destruction of clouds, cannot yet be directly measured. Approximate calculations of the vertical motion are now performed mainly on the basis of theoretical assumptions on the variations of the pressure and wind fields at various heights.

The study of atmospheric fronts and their cloud systems has attracted the interest of many scientists both in the USSR and abroad. Often at the risk of their lives, they fly into thunderstorm clouds and step by step enlarge our knowledge of frontal activity. Front models, worked out mainly by Norwegian* meteorologists (T. Bergeron, S. Petersen and others) have been reconsidered and improved by Soviet scientists. Thanks to the work of A.F. Dyubyuk, N.L. Taborovskii, E.G. Zak, E.K. Fedorov, G.D. Zubyan, E.S. Selezneva and others our knowledge on the formation and occlusion of

* [Sic. Swedish is meant.]

fronts, on the character of vertical air motion and cloud formation, as well as on other problems related to fronts, has been considerably enriched. Yet many important features of cloud formation and of the variation of cloud forms in the evolution of fronts, still remain unknown.

There is a great diversity of opinion on the problem of the vertical extension of fronts in the troposphere and on front formation in the stratosphere. However, more and more scientists have recently arrived at the conclusion that tropospheric fronts in most cases reach the tropopause; they also exist higher – in the stratosphere, but owing to the negligibly small moisture content of the air there, clouds are not formed on stratospheric fronts (G.D. Zubyan, R. Bergern).

Fronts and the weather

In cases where the cloud system of a warm front is similar to the one shown in Figure 54, when the front approaches the observer he will at first see fleecy clouds in the form of hooks, which will be followed by fleecy-strati clouds. As the surface frontline approaches, the air pressure gradually drops. Together with the pressure drop, the temperature and specific

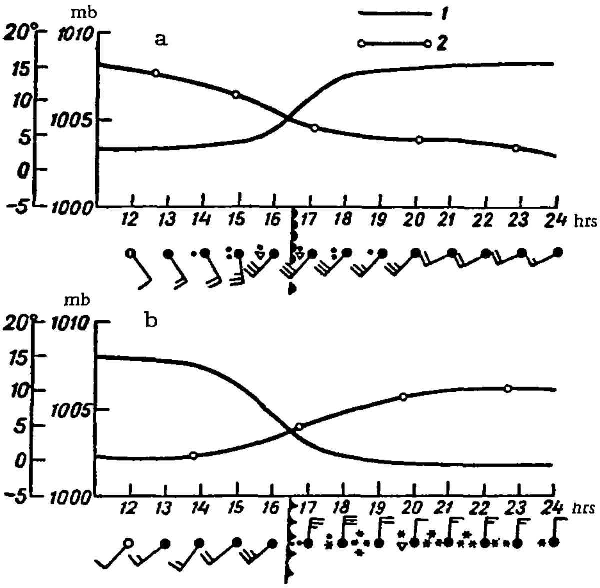

FIGURE 59. Weather variations upon the passage of warm (a) and cold (b) fronts across an observation point

1 – temperature; 2 – pressure.

humidity of the air slowly rise. The clouds become much denser and continuous precipitation begins to fall; this becomes weaker or altogether stops only after the front has passed. During the passage of the front the temperature and specific humidity usually rise sharply, and the wind intensifies. When the front has passed, the wind direction changes and its velocity decreases; the pressure gradually returns to normal and may even begin to rise slightly (Figure 59 a). During winter when a warm front passes, low stratus clouds and fog usually appear.

When a cold front approaches the observation point and the warm air is slightly unstable, cumulonimbus clouds with torrential precipitation appear at first; then, after the passage of the front, stratus-rain and high-stratus clouds appear, and the torrential precipitation is replaced by continuous precipitation which stops as the high-stratus and fleecy clouds approach. As a rule, the width of the precipitation zone of a cold front is smaller than that of a warm front. When a cold front approaches, the air pressure drops, and rises again sharply after its passage. The temperature and specific humidity drop sharply after the passage of the front. The wind before the front may intensify to gale force, and then slowly decrease and change its direction.

Figure 59 b shows the variation of the meteorological conditions upon the passage of a cold front.

CYCLONES AND ANTICYCLONES

Cyclones and anticyclones play an important role in the general circulation of the atmosphere, bringing about interlatitudinal transport of warm and cold air masses. Arising in middle and high latitudes of both hemispheres, cyclones and anticyclones are powerful atmospheric vortexes with horizontal cross sections of 1000–2000 km and more, and of various vertical extensions. In cyclones the atmospheric pressure increases from the center to the periphery. The air currents at the surface of the Earth are directed from the periphery to the center anticlockwise in the northern hemisphere, and clockwise in the southern hemisphere. In anticyclones, on the other hand, the atmospheric pressure increases from the periphery to the center, and the air currents are directed from the center to the periphery clockwise in the northern hemisphere, and anticlockwise in the southern hemisphere. Depending on the intensity and stage of development, cyclonic and anticyclonic motion either extends over the whole troposphere and even the lower stratosphere up to heights of 20–30 km, or are confined to the lowest layers of the troposphere up to heights of 2–3 km.

The appearance and development of cyclones and anticyclones are caused by the unsteadiness of the atmospheric motion. The unsteadiness, i.e., variability of the air currents is determined by many factors, important among which are the nonuniformity of the terrestrial surface, friction, the variation of the horizontal temperature and pressure gradients along the current direction, the curvilinear form of the isobars, etc.

In developing cyclones the air ascends, is cooled, the water vapor contained in it condenses, thick clouds are formed, and precipitation falls. The air flowing at the surface of the Earth into the cyclonic system is thrown high up outside the system thus diminishing the air mass in the central part of the cyclone and correspondingly reducing the atmospheric pressure, i.e., the cyclone becomes deeper.

In developing anticyclones the air descends, is heated, removed from the state of saturation by water vapor, and clouds, as a rule, scatter. Clear or little-cloudy weather prevails. The outflow of air at the surface of the Earth from the system of an anticyclone is compensated by the air inflow at heights, which increases the air mass in the system of the anticyclone and correspondingly increases the atmospheric pressure, i.e., the anticyclone intensifies.

In a system of developing cyclones and anticyclones the mean velocity of ascent or descent of the air is 3–5 cm/sec, or 1–3 m/min. At the same time the mean horizontal velocities of the air currents in the systems of these barometric formations reach 500–1000 m/min.

The pressure at the center of cyclones and anticyclones varies continuously.

From the moment of appearance up to the stage of maximum development the pressure at the center of the cyclone decreases. The increase in the horizontal temperature and pressure gradient in the systems of cyclones and the corresponding intensification of the wind, which often reaches gale force, is connected with this development period. Atmospheric fronts intensify, and cloud formation and precipitation are most intensive. During the winter snowfalls are associated with snowstorms (Figure 60). Afterwards the pressure at the center begins to rise, the winds become weaker, the fronts become occluded, and the precipitation decreases sharply and stops. The cyclone usually fuses with other more powerful cyclones and stops being an independent formation.

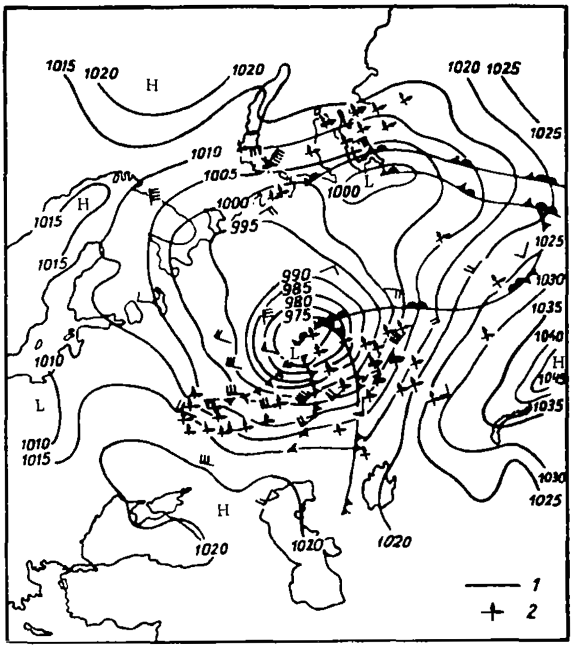

FIGURE 60. Cyclone causing snowstorms over the European Territory of the USSR

1 – isobars; 2 – snowstorms. The arrows show the direction, and the feathering — the velocity of the wind.

In anticyclones, on the other hand, from the moment of appearance until the stage of maximum development the pressure rises at the center. The horizontal pressure gradient and correspondingly also the wind velocities outside the central part increase, the clouds scatter, and clear weather begins. In the second half of the life of an anticyclone the pressure at the center begins to fall and the winds become weaker, usually up to a complete calm. When an anticyclone disappears clouds often appear, and in individual sections precipitation begins to fall.

The pressure at the center of cyclones developing over Europe is often 990–1000 mb. In individual cases they become so deep that the pressure at the center drops to 950 mb and less. This happens most often in the north of the Atlantic and the Pacific oceans. In these cases the wind acquires a destructive force. During the winter over the northern part of America and particularly of Asia cyclones develop on the background of increased pressure. Cyclones with a pressure at the center of 1010–1020 mb are therefore frequent there.

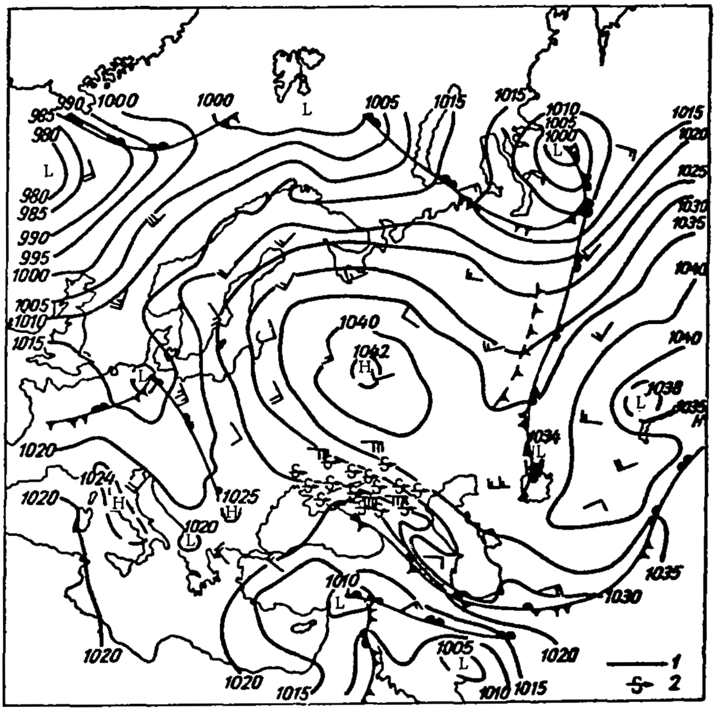

FIGURE 61. Anticyclone causing strong easterly winds (dust storms and sukhovei) over the North Caucasus and south of the Ukraine

1 - isobars; 2 – dust storms. The arrows indicate the direction, and the feathering — the velocity of the wind.

At the center of developing anticyclones the pressure reaches 1030–1040 mb (Figure 61); however, in individual cases it exceeds 1060 and even 1070 mb. Such anticylones arise mainly over the Asian continent during the winter.

How cyclones and anticyclones arise

How do vortexes of such immense dimensions as cyclones and anticyclones arise in the atmosphere?

Until the beginning of the 20th century meteorologists assumed that cyclones were formed as the result of the heating of the air over a warm underlying surface, and anticyclones — as a result of the cooling of the air over a cold underlying surface. This assumption was based on the fact that according to climatological data the atmospheric pressure at middle latitudes in the cold and warm half year is distributed differently over continents and oceans. In fact, if we turn to the maps of the monthly mean pressure for January and July (see Figures 30 and 31), we see that over the cold continents of the northern hemisphere the pressure during winter is high, and over the warm waters of the Atlantic and of the Pacific oceans it is low. During the summer the picture changes radically. High pressure prevails over the relatively cold ocean water, and low pressure — over the heated continents. Moreover, even over such small but warm water basins as the Mediterranean and Black seas, reduced-pressure regions appear during winter; these owe their origin to the temperature difference between land and sea.

This hypothesis was rejected at the very beginning of the 20th century, on the basis of aerological data collected by means of sounding balloons, and observational data in mountains. Observations showed that the temperature in the whole troposphere in a system of cyclones is on the average much lower than in a system of anticyclones. In Europe the temperature difference between these barometric formations in the layer from the surface of the Earth to a height of 9 km is equal on the average to 5.4°C. It is true that over cyclones the stratosphere is warmer than over anticyclones. The temperature difference in the layer 9–16 km is on the average 5.5°C. But this cannot explain the essence of the thermal generation theory of cyclones and anticyclones.

During the 1920's the Norwegian meteorologists made a hypothesis of wave character of the development of cyclones. They assumed that cyclones appear as a result of wave (oscillatory) motions of interfaces existing in the atmosphere between air masses of different densities (frontal surfaces). A theoretical solution of the problem of the generation of a cyclonic disturbance was proposed by N.E. Kochin. The wave theory has, however, a big shortcoming: it does not give indications on the subsequent development of cyclones. Moreover, it does not relate the appearance and development of cyclones with the causes of variation of the atmospheric pressure. In fact, this problem is not considered at all. As a result, anticyclones were not included in the wave theory.

New opinions on the nature of the appearance and intensification of barometric formations, in particular of anticyclones, were proposed in 1932 by V.M. Mikhel. On the basis of a study of the evolution of anticyclones he established that anticyclones intensify in cases where the air currents at heights of 5–6 km converge and, conversely, anticyclones become weaker where the air currents diverge. In addition V.M. Mikhel and S.I. Troitskii established that anticyclones (as well as cyclones) move in the direction of the main "leading" current in the middle troposphere.

In Germany during the thirties, on the basis of the assumptions of V.M. Mikhel, R. Scherhag put forward the so-called divergence theory of cyclones. This states that air masses are thrown aside because of the divergence of the air currents at heights. As a result of the decrease in the air mass the atmospheric pressure drops and a cyclone develops.

At the end of the thirties N.L. Taborovskii and the author put forward an advective-dynamic theory of appearance and development of cyclones and anticyclones based on an analysis of the advective and dynamic variations of the temperature and pressure and of the transformation conditions of high-level deformation fields. According to this theory, the appearance and development both of cyclones and of anticyclones is organically related to the variation of the atmospheric pressure. As a result of the horizontal transport of cold and warm air masses, i.e., of temperature advection, high-level frontal zones with a definite structure of the air currents appear and intensify in the troposphere. The arising unsteady motion causes the throwing out or squeezing of the air and consequently, also the decrease or increase in the atmospheric pressure. These pressure variations are called "dynamic pressure variations." Under a prolonged temperature advection and intensive dynamic pressure variation, cyclones or anticyclones appear and then develop, depending on whether the air mass in the system of these barometric formations decreases or increases. Quantitative criteria for the temperature contrasts and wind velocities in the troposphere, necessary for the development of cyclones, were also established.

However, the advective – dynamic theory of cyclones and anticyclones still only takes into account a few of the factors causing modification of the atmospheric pressure.

The difficulty of creating a rigorous quantitative theory of the appearance and development of cyclones and anticyclones is due to the fact that even now there is no theory giving a complete description of the complex process of pressure variations.

It is known, of course, that the atmospheric pressure is determined by the weight of the air column, the variation of the weight depending both on the variation of the air density, which may result from temperature variation, and on processes leading to an increase or decrease in the air mass. Thus thermal and dynamic factors should be taken into account when determining the pressure variation and the appearance and development of cyclones and anticyclones.

The influence of the thermal factors is closely related to the transport of air masses of different densities, causing variations in the horizontal temperature, pressure, and wind fields. But these fields also undergo continuous variations as a result of the vertical motion of the air, the heat inflow from the underlying surface, the heat losses on evaporation, the release of latent heat, and so on.

Great difficulties are encountered when trying to arrive at a quantitative estimate for the above listed processes. These are caused mainly by the absence of reliable observational data on the air humidity, turbulence, vertical motion in various layers of the troposphere, etc. More accurate data are also needed on the distribution and variation of the temperature above 15–20 km. Therefore, when estimating the influence of the temperature on the variation of the pressure field, one has to settle for approximate and average data.

The absence of a quantitative theory satisfactorily describing the complex processes of pressure variation, and the appearance and development of cyclones and anticyclones is a result of our lack of knowledge of the role of various factors in the variation of the atmospheric pressure. Thanks to the data of aerological observations, we now have clearer ideas about the structural features of cyclones and anticyclones. They appear as a result of a drop or rise in the atmospheric pressure at the surface of the Earth in the form of vortexes of small dimensions (with a diameter of up to 400–600 km). As they develop they gradually expand, drawing in huge masses of air.

Three stages in the development of cyclones and anticyclones are usually defined: appearance, maximum development, and decay and disappearance.

In the first, initial stage, small barometric disturbances occur, outlined by one or two isobars, with a pressure difference between the center and the periphery of up to 5–10 mb and with a definite wind system at the surface of the Earth. At heights of 2–3 km, closed isobars are not observed.

The second stage is the stage of maximum development of the barometric formation with minimum pressure at the center of a cyclone and maximum pressure at the center of an anticyclone. The pressure difference between the center and the periphery often reaches 20–30 mb and more. In this stage the corresponding circulation system is observed in the upper troposphere at heights of 8–12 km.

In the third stage – the decay stage – the concentric system of isobars is traced not only at the surface of the Earth but also at heights. In these cases the cyclonic circulation usually extends not only throughout the entire troposphere, but also in the lower layers of the stratosphere. However, cyclones gradually fill up.

Spatial structure of cyclones

One may get an idea of the spatial structure of a developing cyclone from a diagram containing two isoline systems. In Figure 62 the thick solid lines represent the isobars, i.e., the lines of equal pressure, at the surface of the Earth; the thin lines represent the isobars at a height of about 6–8 km in the horizontal plane. From the latter it is possible to get an idea of the direction of the air currents at heights, since at these heights the direction of the wind is approximately that of the isobars.

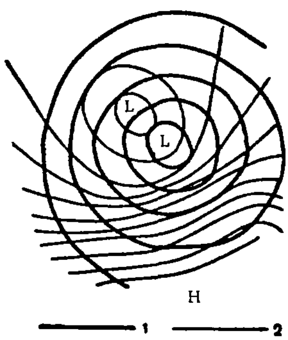

FIGURE 62. The pressure field at the surface of the Earth (1) and at heights (2) in the system of a developing cyclone

The energy reserves of a cyclone can be qualitatively determined from the density of the isotherms. In the initial stage of development of a cyclone (after its appearance) the temperature contrasts in the system are quite large, amounting to 8–10°C above 1000 km (Figure 63 a). In a deepening cyclone the temperature contrasts as well as the density of the isotherms increase (Figure 63 b). Afterwards the temperature contrasts decrease and the cyclone, devoid of energy reserves, gradually fills up with cold air in the whole troposphere (Figure 63 c).

Thus, over the period of the deepening of a cyclone which appeared at the surface of the Earth, large horizontal temperature and pressure gradients were observed in the troposphere. The wind velocities increased with height, reaching in the middle and upper troposphere 100–200 km/hr and more.

The cyclone moves in accordance with the structural features of the temperature and pressure fields and the structure of the air currents determined by them, its motion taking place along the air current prevailing in the troposphere. In our case the cyclone can move from left to right in the plane of the figure. The temperature distribution in the troposphere in the system of a cyclone is such that if we look in the direction of motion, the cold air is on the left and the warm air – on the right.

A cold front is always situated in the tail section of a moving cyclone and a warm front — in the front section. In the system of a cyclone the isotherms are as though carried along the air currents. On the left side of the cyclone air is transported from low to high temperatures, and on the right, i.e., in the front section, the air moves from high to lower temperatures. Therefore, when a cyclone approaches a given point the temperature *first rises and then drops appreciably.*

During the period of deepening of a cyclone, i.e., when the pressure at the center is decreasing, the number of closed isobars increases, and the wind and the atmospheric fronts in their systems intensify. Intensive vertical air ascent, intensified cloud formation, and fall of precipitation are

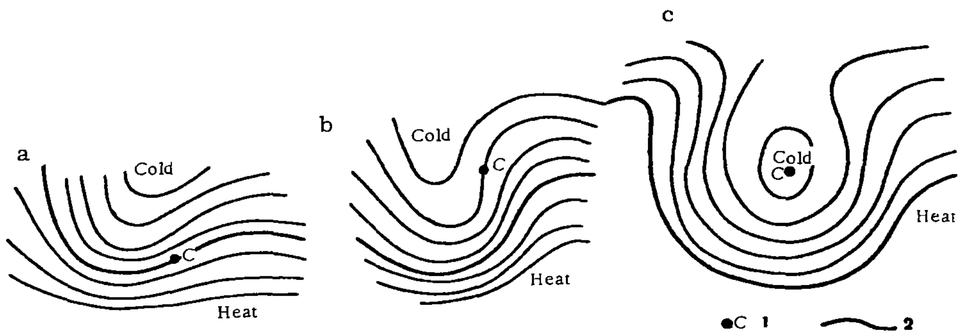

FIGURE 63. The temperature field in the system of a cyclone in various stages of development

a – appearance; b – maximum depression; c – filling-up; 1 – surface cyclone center; 2 – separation line between warm and cold air masses.

observed. These processes are particularly well displayed in the front section of a cyclone, in the zone of the warm front and at the center. In the tail section of a cyclone the air ascent is weak, and air even descends on the periphery, scattering the clouds and stopping the precipitation. In most cases, therefore, a pressure drop indicates the approach of a cyclone and the worsening of the weather, and a pressure rise indicates that a cyclone moves further away and that the weather improves.

However, this rule has exceptions, since the weather is not always worse at the central part of a cyclone with its low pressure than on the periphery of a cyclone and even on the periphery of an approaching anticyclone. In the latter case, in spite of the rise in the atmospheric pressure, the weather may worsen. This depends on the peculiarities of the processes taking place in the cyclonic system, and mainly on the relationship between the velocity of the ascending motion and the moisture content of the air. When the moisture content is low and the air is far from the state of saturation, the condensation process is weak and precipitation may not fall.

It should be noted that as a cyclone develops new air masses are drawn into its system. During the life of a cyclone the air masses are replaced two to three times, carrying with them water vapor, which, when condensing and falling in the form of precipitation, irrigates a huge territory over which the cyclone moves.

It has been established that the deepening of a cyclone takes place as long as *colder air masses penetrate to its system from the outside through the* tail section. As soon as the inflow of cold air gets weaker, and all the more, stops, the cyclone begins to fill up, i.e., the pressure drop at the center is replaced by a pressure rise.

The cooling of the air in a cyclonic system and the filling up of this barometric formation by homogeneous masses of cold air are connected in the last stage of development not only with the inflow of cold air from the outside, but also with the adiabatic cooling, caused by ascending motion of the air.

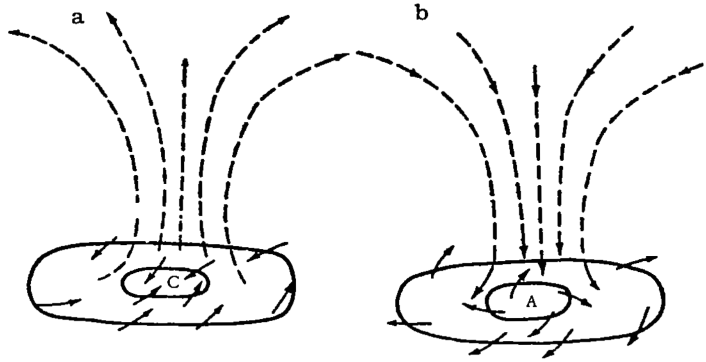

FIGURE 64. Schematic diagram of the vertical currents in a cyclonic (a) and an anticyclonic (b) system

The axis of a cyclone is strongly inclined toward the cold air as long as temperature contrasts exist in its system. When the cyclone is completely filled up with cold air, the cyclone's axis levels off and can be considered directed along the vertical.

As a cyclone deepens and fills up with cold air the tropopause over it drops. Over cold filling-up cyclones, therefore, the tropopause lies lower than over warm deepening ones.

The processes developing in the troposphere and in the lower stratosphere are closely related. A cooling of the air in the troposphere is associated with a temperature rise in the lower stratosphere. In this process, besides advection, a large role is played by descending motion and adiabatic heating of the air in the lower stratosphere. And conversely, as the air temperature rises in the troposphere (in developing anticyclones) the tropopause ascends and the air temperature in the lower stratosphere drops considerably.

The system of vertical currents in cyclones and anticyclones is shown schematically in Figures 64 a and b. In the first system, air flows in near the surface of the Earth (shown by small arrows), moves upward, and flows out of the system; in the second case, on the other hand, the air flows in at the upper layers, descends, and flows out of the system near the surface of the Earth.

Spatial structure of anticyclones

Anticyclones appear, develop, and decay in the same way as cyclones, but with the important difference that under the action of the forces in the system of a developing anticyclone air masses accumulate, bringing about a rise in the atmospheric pressure, a descending motion, and a divergence of the surface wind. As a consequence, atmospheric fronts in the system of such an anticyclone are eroded at the surface of the Earth and clear weather usually prevails.

Figure 65 shows an anticyclone with the isobars at the surface of the Earth and higher up (in the horizontal plane) in the second state of development.

A surface anticyclone moves in accordance with the direction of the current prevailing higher up. It carries in its front section cold, and in its tail section – heat. As the anticyclone develops, the fronts are eroded. The vertical thickness of an anticyclone varies, depending on the stage of development.

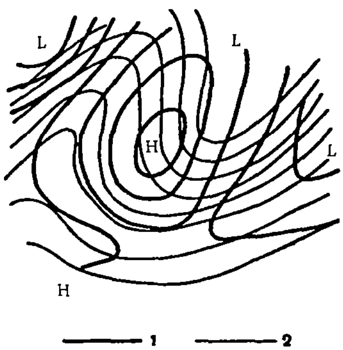

FIGURE 65. The pressure field of an anticyclonic system in the second stage of its development at the surface of the Earth (1) and in the upper troposphere (2)

At the beginning of the development of an anticyclone the anticyclonic circulation extends only over the lower air layers, up to 2–3 km. As it expands and intensifies, the anticyclonic circulation extends over ever larger air masses. In the last stage of development, an anticyclone often reaches the lower layers of the stratosphere. During the period of their intensification, the temperature contrasts in anticyclones as well as in cyclones increase in the whole troposphere. After this, as warm air masses are drawn into the system of the anticyclone, the horizontal thermal asymmetry is nullified, and the anticyclone becomes a barometric formation with almost homogeneous warm air. The adiabatic heating of the air resulting from its descent, which is characteristic of anticyclones in the period of their intensification, contributes to this.

With the vanishing of the horizontal temperature contrast, anticyclones begin to collapse. The vertical axis of an anticyclone in the presence of horizontal temperature contrasts is inclined toward the warm air. As the anticyclone is filled up with masses of warm air and the thermal asymmetry vanishes in its central part, the axis tends to a vertical position.

High cold cyclones and warm anticyclones become new focuses of cold and heat in the troposphere. On the periphery of these filling-up cyclones and collapsing anticyclones new tropospheric frontal zones are formed with large temperature contrasts and strong air currents. New atmospheric fronts and new cyclonic and anticyclonic disturbances arise, which under appropriate conditions pass through a full cycle of life.

Atmospheric fronts and systems of cyclones and anticyclones

In addition to the main fronts, cyclones often contain secondary fronts caused by new portions of cold air entering the system of the cyclone through its tail section and by the wind convergence in the surface air layer. During the cold time of the year, when the surface air layer over the continent cools down, the cold air in the tail is stable and, as a rule, clouds are not formed. During the warm time of the year, instability and the accompanying convective cloudiness develop due to the heating of the lower air layers, which in the case of sufficient moisture content of the air cause torrential precipitation.

Due to the difference in the velocities of the cold and warm fronts they join and the cyclone is occluded. As the air cools down in the system of the cyclone, the fronts move to the periphery. Becoming thermally uniform and cold, the cyclone slows down until it almost stops, i.e., is transformed into a stationary cyclone. In this last stage, the cloudiness is eroded and cloudless layers appear. During the winter weak precipitation falls, and during the summer, as a result of the heating and of the development of instability, cumulonimbus clouds are formed and torrential rains fall.

In anticyclones the fronts behave differently. In accordance with the air currents at the surface of the Earth, they move to the periphery of the anticyclone and are eroded. However, in the intensification period of the anticyclone, the temperature contrasts in the troposphere are maintained. Since clear weather prevails in anticyclones, the heating or cooling of the terrestrial surface and of the adjacent air layers is more intense than in cyclones, where the multilayered cloudiness diffuses the solar short-wave radiation. In clear weather the heating or cooling of the lower air layers in anticyclones is most rapid.

Over the continents during winter, a temperature inversion arises as a result of the air cooling. As was shown above, in Siberia at heights of 1.5–2.0 km the winter air temperature is often higher than at the surface of the Earth by 10°–20°C and more. Two air layers are formed: cold below, and relatively warm above. If the moisture content of the air is low, clouds are not formed at the lower boundary of the inversion layer; this is often observed in Siberia at low temperatures. Over the European territory of the USSR the under-inversion air usually contains sufficient amounts of moisture. As a result under-inversion clouds of stratus and straticumulus forms of small vertical thickness appear. Inversions are most often formed at a height of about 1 km.

Over the continents during summer on the other hand the lapse rates rapidly increase owing to the heating of the lower air layers, and convection appears. If the moisture content of the air is sufficient, cumulus and cumulonimbus clouds are formed, and downpours and thunderstorms begin. The prediction of torrential precipitation at a given point is very difficult. One of the many reasons for this is that torrential precipitation does not fall everywhere. For example, torrential precipitation often falls over particular places in Moscow but not over the entire town. The precipitation forecasting is therefore borne out only for those regions of the town where the precipitation fell.

Tropical cyclones

Until now we have dealt only with extratropical cyclones, i.e., with cyclones born at extratropical latitudes.

Cyclonic vortexes are, however, formed in the tropical regions as well. In eastern Asia, they are called typhoons; in Chinese this means "great wind."

The dimensions of tropical cyclones are relatively small, having diameters of from tens to hundreds of kilometers. The pressure gradient in tropical cyclones reaches 20–40 mb per 100 km, and the wind velocity usually exceeds 100–150 km/hr.

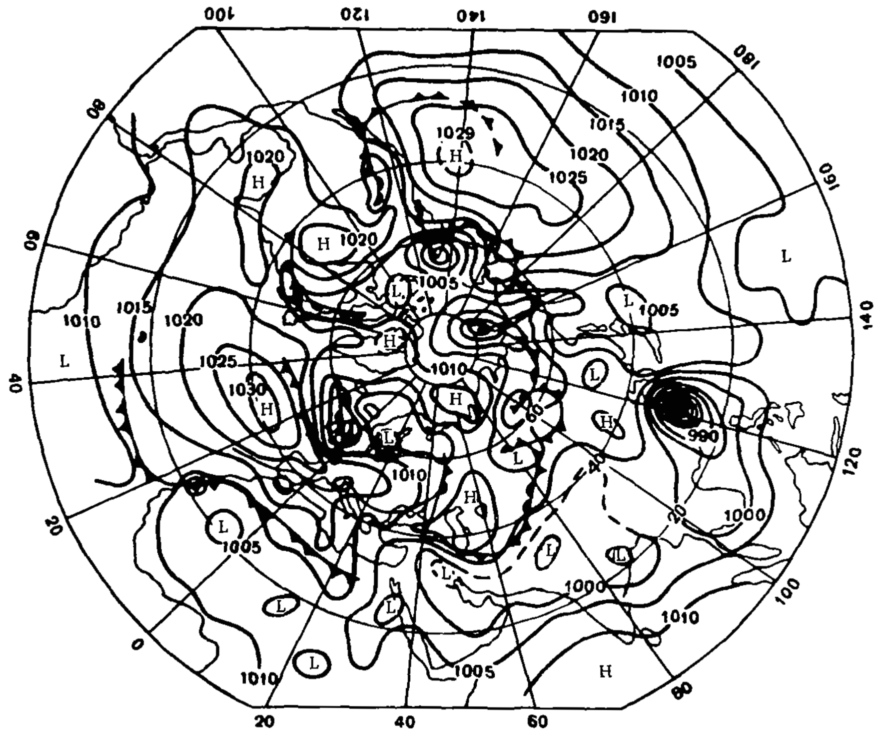

FIGURE 66. The pressure field in the northern hemisphere at ground level on the evening of 31 July, 1956. Typhoon located in southeast Asia

On the surface pressure map of the northern hemisphere for the evening of 31 July, 1956 (Figure 66), in addition to the other barometric formations, a tropical cyclone is clearly shown over southeast Asia by the density of the isobars there. It arose over the Pacific Ocean at the latitude 15° and in 2–3 days reached the shores of Asia, where it filled up.

Tropical cyclones are born in the calm zone over the oceans (mainly between the latitudes 10° and 20°) both in the southern and in the northern hemispheres. They then move along the isobars from east to west. In the northern hemisphere the tropical cyclones appearing over the Pacific Ocean reach the southeastern shores of Asia, and then turn to the right and move toward the Japanese islands. The average number of typhoons born annually at the southeastern shores of Asia is over 20. Over the Atlantic, tropical cyclones also move along the trade winds. Reaching the Gulf of Mexico and Florida they turn to the north. At the middle latitudes, tropical cyclones are as though regenerated (reborn), transforming into powerful cyclones with a well-pronounced temperature asymmetry typical of the cyclones of the extratropical latitudes. Tropical cyclones are often observed in Indo-China, on the Pacific Ocean coast of China and in Japan. In individual cases they appear in the Far East territories of the USSR and on the Atlantic Coast of North America. Less often tropical cyclones are formed over the north of the Indian Ocean (Figure 67).

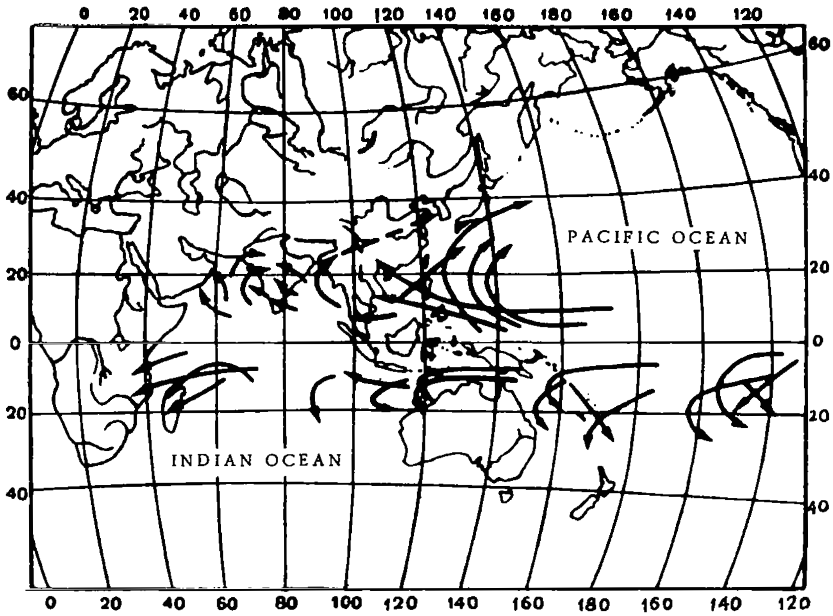

FIGURE 67. Trajectories of tropical cyclones over the Pacific and Indian oceans

In the southern hemisphere tropical cyclones appear in the equatorial zone of the Indian and Pacific oceans. They do not appear over the South Atlantic. The circulation system in tropical cyclones is similar to the circulation in the cyclones of extratropical latitudes: in the northern hemisphere anticlockwise, in the southern hemisphere, clockwise.

The reasons for the appearance of typhoons and extratropical cyclones are not the same. Whereas for the appearance of cyclones of the extratropical latitudes large horizontal temperature and pressure gradients in the troposphere are necessary, in the beginning of tropical cyclones they almost do not exist. In the system of tropical cyclones there are no atmospheric fronts, with the exception of cases of their regeneration and transformation into ordinary cyclones of middle and high latitudes.

The reasons for the appearance of tropical cyclones are as yet little known. It is assumed that their formation is due to a large thermal instability of the air when it is sufficiently moistened. In the northern hemisphere tropical cyclones are formed mainly during the second half of the summer and during the fall in the calm zone. During spring and the first half of the summer their appearance is rare and in January–April there are none. August, September, and October are the months with the most frequent formation of tropical cyclones. In the southern hemisphere over the Indian and Pacific oceans, they most often appear in December–March, and only in exceptional cases in May–October.

The velocity of tropical cyclones is considerably lower than that of cyclones of middle and high latitudes. At low latitudes their velocity rarely exceeds 15–20 km/hr, or 350–500 km/24 hrs, i.e., it corresponds to the trade wind velocities. A very interesting phenomenon is observed in tropical cyclones. Bad weather with hurricane winds is characteristic of all their system with the exception of the center, where a section of blue sky, the so-called "storm's eye" is usually seen (Figure 68). This is connected with the existence of descending air motion at the center, whereas in the whole system of tropical cyclones the air moves intensively upward.

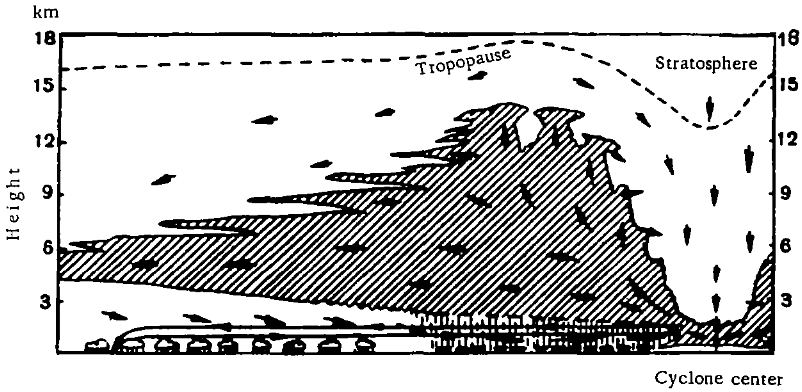

FIGURE 68. Vertical cross section through the center and left-hand side of a tropical cyclone at the southeastern shores of North America

The dashed section represents the cloudiness, the arrows give the direction of the air current.

The destructive force of tropical cyclones is tremendous. The wind velocity often reaches 300–400 km/hr. For example, the typhoon which passed over Japan on 21 November, 1934 destroyed partially or completely up to 700,000 houses, put out of action over 11,000 vessels, and caused floods and tremendous damage.

Another such typhoon was the one which rushed past Japan on 26 September, 1959. According to newspaper reports the wind velocity reached 180 km/hr. Such a wind tears roofs from houses, uproots trees, and destroys everything in its path. Gusts of wind, currents of rain, and sea waves caused destruction in many towns and villages. Up to 1.5 million persons remained homeless. Over 5000 persons died or were reported missing, over 15,000 were wounded; 180,000 houses were destroyed, and about 300,000 homes were flooded. Railroad transport, sea vessels, etc., suffered.

According to "Pravda" of 20 September, * tropical cyclones caused frightful destruction on the coasts of the Atlantic and Pacific oceans during the summer of 1961. One of them, "Carla" moved on 6 September from the Gulf of Mexico to the states of Texas and Louisiana. The town Gulfport,

* See the paper: Batyaeva, T. and L. Minina. Vetry tropicheskogo poyasa (Winds of the Tropical Belt).

on the coast of this gulf, was almost completely destroyed. The wind of over 200 km/hr carried light wooden structures with it. Another tropical cyclone ("Debby"), appearing near the islands off Cape Verde, moved to the British Isles, where it caused great destruction and subsequently filled up in the region of the Norwegian Sea.

Even more destructive cyclones appeared over the Pacific Ocean. On 4 September the typhoon "Pamela" appeared in the region of the Marshall Islands; it raged for several days over Taiwan. In Taipei alone 800 houses were destroyed.

Several days later, again near the Marshall Islands, the cyclone "Nancy" appeared with a wind velocity exceeding 300 km/hr. On 15 September it approached the southern coast of Japan and passed along the island to the northeast, destroying along its path over 450,000 houses, and 400 bridges and dams. According to incomplete data, over 150 persons perished and over 2000 persons were wounded. In many regions the railroad transport was interrupted, and electricity supplies cut. The passage of the typhoon "Nancy" was accompanied by strong torrential rains, and the coastal regions were inundated by oceanic waves. On 17 September the typhoon arrived at the Sea of Okhotsk and caused destruction in the southern part of Sakhalin Island.

The tropical cyclone "Nancy" was one of the strongest typhoons of recent years.

Typhoons with a somewhat smaller destructive force are observed every year. They sometimes cause damage to populated regions in the Far East territories of the USSR when they move somewhat to the west of their usual path.

Regions of cyclogenesis

Huge moving atmospheric vortexes — extratropical cyclones and anticyclones — appear and develop in particular regions more often than in others. In some regions cyclones appear more often, in others — anticyclones. Depending on the frequency of the barometric formations, there are various types of weather and climate. In places where cyclones often appear and develop, overcast or cloudy weather prevails. The amount of precipitation almost always exceeds the amount of evaporating moisture, and therefore the runoff in these regions reaches large mangitudes.

In regions where anticyclones appear and develop more often, clear weather prevails. The amount of atmospheric precipitation is insufficient, and the related runoff is small or altogether nonexistent. Closely connected with the climatic conditions, regions of prevailance of cyclonic weather are rich in vegetation, whereas in regions of anticyclonic weather the vegetation is quite scanty. There, arid steppes and deserts are formed.

Frequency maps of cyclones, prepared by the author for the northern hemisphere, show that cyclones appear and develop most often at the eastern coasts of North America and Asia during the winter. Over the Atlantic they move in the direction of Iceland and the Barents Sea and over the Pacific Ocean, from Japan to the Aleutian Islands and Alaska. The cyclones are usually most powerful and deep over the regions of Newfoundland and Iceland as well as over Japan and the Aleutian Islands. Their activity there causes the fall of considerable precipitation, strong winds, and frequent gales. Quite often during the winter, cyclones also appear over the Mediterranean Sea.

The frequency of cyclones is considerably lower over the continents than over the oceans; the appearance of cyclones over the Asian continent is particularly rare during winter.

Whereas during the winter strong snowfalls and snowstorms to the north of the European territory of the USSR are very often caused by cyclones arriving from the north of the Atlantic, the considerable precipitation to the south of the USSR is usually connected with cyclones arriving from the direction of the Mediterranean Sea. However, moving in a northerly or northeasterly direction, cyclones often cover tremendous distances and give a considerable amount of precipitation, and in places snowstorms, over the territory lying along their path.

During the summer the situation is somewhat different:

1) The frequency of cyclones decreases over the oceans and increases over the continents.

2). The southern boundary of the formation zone of cyclones shifts to the north. Whereas during winter they appear at latitudes down to 30°, and sometimes even more to the south, during summer the formation zone of cyclones is bordered on the south approximately by the 40th parallel (in the northern hemisphere). In particular, Mediterranean Sea cyclones appear very rarely during the summer.

3) The intensity of summer cyclones is considerably lower than winter ones.

Owing to the shift of the cyclone zone to the north during summer the precipitation in the subtropics decreases or even temporarily stops. For this reason, in the south of Italy, the Mediterranean Sea, North Africa, and the Near East, atmospheric precipitation falls mainly during the winter half year. Rains during the summer months are rare in these regions.

In the southern hemisphere extratropical cyclones most often appear at middle latitudes. As in the northern hemisphere they mostly move from west to east.

Regions of cyclonic and anticyclonic activity

During January in the northern hemisphere the highest frequency of moving anticyclones is recorded on the continents of Asia and North America. It is characteristic that the maximum number of January anticyclones is observed not only over these continents, but also over the adjacent parts of the oceans. The lowest frequency of anticyclones is, as a rule, in the regions where the highest frequency of cyclones is observed, and vice versa.

Cyclones and anticyclones move along the prevailing high-level air currents. Thus, over the Atlantic they move mainly from the region of the eastern coast of Africa to the regions of Iceland and the Barents Sea.

A similar situation can be observed in the Pacific Ocean. The main paths of cyclones there also coincide with the direction of the prevailing air currents. The maximum frequency of cyclones is recorded approximately in the region of 45° NL and 170° E. long. Cyclones appear to the west of this region (up to the coast of Asia), and to the east of this region, cyclone filling up prevails.

Anticyclones also move along the high-level currents over the continents. Thus, over the European territory of the USSR, in accordance with the structure of the currents at high levels during winter, anticyclones move from west to northwest. Forming and intensifying over the continent, they continue their motion to the east. The same occurs over eastern Siberia, the Far East, and North America.

In the chapter dealing with air currents and atmospheric fronts, we mentioned that as a result of the nonuniform inflow of heat from the Sun and of the interlatitudinal exchange of air masses, high-level frontal zones characterized by large horizontal temperature and pressure gradients and by quite strong winds are formed. As a rule cyclones and anticyclones appear and develop under frontal zones. Although these zones in the extra-tropical latitudes continuously form and disappear everywhere, there are regions where they appear and exist more frequently than others. On the maps of the barometric topography, given in the chapter on the air temperature and pressure, one may clearly see high-level frontal zones surrounding the northern and southern hemispheres.

The regions of maximum frequency of cyclones and anticyclones territorially coincide with the zones of maximum density of the temperature and pressure isolines.

All that was told in this chapter about the appearance and movement of barometric formations during winter also refers to other seasons of the year, in particular to the summer, with the only difference that summer atmospheric processes are less intense. This, as was said above, is due to the general decrease in the horizontal temperature and pressure gradients, as well as to the decrease in the velocities of the air currents during the summer.

Cyclonic and anticyclonic activity exists during the summer also in those places where, owing to the interlatitudinal exchange of air masses, frontal zones with temperature contrasts up to 10°–15°C per 1000 km and more with considerable velocities of the air currents arise and intensify.

However, the intensity of cyclone activity is lower during the summer. Then the zone of maximum frequency of cyclones over the north of the Atlantic coincides basically with the regions of maximum isotherm density on the relative-topography map for July. In contrast to the winter, during the summer over Siberia, in the region of increased temperature contrasts, in the headwaters of the Ob and Enisei, the frequency of cyclones increases. Appearing every now and then, they move toward the east and mature over eastern Siberia and the Far East.

Due to some weakening of the interlatitudinal circulation during summer, *cyclones move more frequently from west to east along the parallels.*

Moving anticyclones form most often over the oceans, the Mediterranean Sea, eastern Europe, western Siberia, and the eastern regions of North America. During the summer, stationary anticyclones are observed more often over the Atlantic and Pacific oceans, between the latitudes 25° and 40°. During January they form most frequently over the west of North America and over eastern Siberia.

From winter to summer the structure of the pressure field at the surface of the Earth undergoes considerably larger variations than higher up. In the Baikal region the monthly mean pressure of July drops by 30 mb as compared with the pressure during January. Due to the low frequency and intensity, the Icelandic and particularly the Aleutian cyclones are very weak on the surface mean-pressure map. Over South Asia there is a low-pressure region throughout the warm half year owing its existence to the heating of the continent and of the air, since the heating of the air at that region is considerably more intense than over the surrounding territory. For precisely this reason, this relatively stationary south Asian depression is called thermal. Such thermal depressions are formed in the summer half year over North Africa and Mexico. Due to the small dimensions of the continents, these depressions are, therefore, not so extensive as the south Asian one.

JET STREAMS IN THE ATMOSPHERE

Strong winds and the troposphere

We discussed the appearance of high-level frontal zones with which the most active atmospheric processes and weather variations are connected, in the chapter dealing with the horizontal nonuniformity of the atmosphere and atmospheric fronts.

The study of the conditions for the formation and decay, as well as of the spatial structure of high-level frontal zones began only in the 1930's. The features of these zones were first studied in the more accessible part of the troposphere, mainly between the surface of the Earth and the 500-mb isobaric surface, i.e., the level of about 5 km (R. Scherhag, M. Rodeval'd, Kh.P. Pogosyan, N.L. Taborovskii and others). This was due to the fact that in the thirties the sounding of the atmosphere at the network of stations was carried out mainly by means of airplanes which reached heights of only 4–6 km. Radio sounding was also not yet perfected and thus not in wide use.

Due to radiosondes and high-flying airplanes, the upper layers of the troposphere and the lower stratosphere became accessible for extensive study. Attention was mainly concentrated on the structural features of the temperature and pressure fields, on determining the horizontal transport (advection) of air masses possessing various properties, on studying the conditions for the appearance and development of barometric formations, as well as of the appearance and decay of atmospheric fronts etc.

During the last ten years much attention has been paid to the study of the wind in the system of high-level frontal zones. From the theory of the thermal wind, as well as from data of pilot-balloon observations, it was known that, in accordance with the temperature distribution with height, the wind velocity usually increases up to the level of the tropopause, and in the lower stratosphere it decreases, i.e., the maximum air-current velocities are observed at the 9–12-km level, near the tropopause.

It should be noted that the first information about wind velocities higher up already appeared at the turn of this century. However, until the development of aviation and the rise in the ceiling of airplanes, strong winds did not attract special attention. During October 1917, for example, a large number of German "Zeppelins," returning from an attack on England, ascended to a height of 6000 m, were dragged into a strong northerly air current, knocked off course, and some were destroyed.

At about this time the existence of wind velocities greater than 200 km/hr in the upper troposphere was also established from the movement of cirrus clouds. However, the data were insufficient for more definite conclusions about the wind velocities in some geographic regions.

First indications on the distribution of the monthly mean air velocities at various heights in various regions of the Earth were obtained by means of pressure maps. Pilot-balloon observations of the wind higher up were few, and it was impossible to determine the characteristic seasonal features of the regime of air currents over various geographic regions in various atmospheric processes.

On the other hand, developing aviation demanded more than a widening of knowledge about the wind regime at various heights. It also required the development of methods for forecasting the wind along flight routes. Whereas at the beginning these requirements referred to the middle troposphere, the interest in the upper troposphere and in the lower stratosphere grew with the rise in the ceiling of airplanes. Gradually the forecasting of the wind high up became one of the principle requirements of weather forecasting for aviation.

The increased interest of pilots in the forecasting of the wind high up is due to the fact that air currents along an aviation route may either assist, or oppose the flight of the airplane. Depending on the airplane's velocity and on the velocity of the wind, the airplane may gain or lose from 0 to 100% of its velocity. Since the wind direction and speed are different at different levels, one can choose the most advantageous level for the flight from the nature of the wind along the flight route at various levels.

During the second world war cases were recorded when bombers flying at great altitudes were considerably deflected from the planned flight course. Moreover, there were great losses to airplane velocities. At that time war planes usually flew at heights of 5–10 km with velocities of 300–400 km/hr. Strong counter winds obviously did not assist the flight. Cases are known of airplanes, flying at heights of 8–10 km with velocities of 300–400 km/hr, falling over the Japanese islands into the zone of a westerly counterwind of such force that, having lost velocity, the airplane remained almost motionless in the air. Shortly afterwards such wind velocities were also observed over other regions of the Earth.

The turbulence of the atmosphere at the level of the high wind velocities plays a large role in the safety of flights. In the zone of maximum wind velocities, airplane bumpiness caused by atmospheric turbulence is often observed. There were cases when bumps were strong enough to destroy the airplane.

Observations showed that air-currents with high velocities are concentrated in a relatively narrow zone in the form of jets situated in a relatively calm surrounding atmosphere. The zones of high wind velocities were therefore called *jet streams*.* The discovery of the jet streams may be considered as one of the greatest meteorological discoveries of the middle of this century.

A jet stream is a strong, narrow current of large extension in the upper troposphere and lower stratosphere having considerable velocity gradients, passing along the line of maximum winds, and having an elliptic vertical cross section. In the system of the jet stream the maximum velocity on the jet axis exceeds 100 km/hr.

Jet streams are about one kilometer deep, several hundreds of kilometers wide, and several thousands of kilometers long.

* For more details see the book: Pogosyan, Kh.P. Struinye techeniya v atmosfere (Jet Streams in the Atmosphere). —Moskva, Gidrometeoizdat. 1960.

The information about the wind regime above the level of the tropopause considerably increased only after the second world war, when radar installations became available.

The tropospheric jet streams

The velocities of air currents higher up depend mainly on the character of the temperature field of the lower lying air layers. The larger the horizontal temperature gradients in the system of a high-level frontal zone, the stronger the jet stream, indicating the existence of strong winds in this zone. In other words, in the formation and evolution of jet streams the main role is played by the temperature distribution in the atmosphere and the arising horizontal temperature gradients.

Jet streams are usually connected with high-level frontal zones. They appear, intensify, or weaken as a result of the appearance and decay of tropospheric fronts. In the first case the horizontal gradients of the temperature, pressure, and wind velocity increase as a result of the approach of cold or warm air masses. In the second case, when the cold and warm air move apart, the temperature and pressure gradients decrease, and the winds become weaker.

Jet streams arise in the troposphere and stratosphere. In the troposphere they are almost constantly observed in the subtropical zone of the northern and southern hemispheres: during the winter between the latitudes 25° and 35°, during the summer between 35° and 45°. Jet streams very often appear and develop in the troposphere at extratropical latitudes up to the Central Arctic and the Antarctic. Depending on the regions in which they appear in the troposphere they are divided into *subtropical* and *extratropical* jet streams.

The highest wind velocities in the troposphere are usually observed near the tropopause. Data on the wind distribution high up show that the highest velocities are most often observed under the tropopause and more rarely above it. In the stratosphere they are observed at heights of 25–30 km during the winter under certain circulation conditions.

Tropospheric jet streams are observed over almost the entire terrestrial globe, but not everywhere with the same frequency. There are, for example, regions where at heights of 9–12 km the maximum velocities in the jet almost always exceed 200 km/hr. This occurs particularly along the Pacific Ocean coast of Asia at the latitude 30°–40°. There, particularly over the southeastern part of China and over the Japanese islands, wind velocities (mainly of a westerly direction) exceeding 200 km/hr are usual at heights of 9–12 km for 6–8 months of the year.

Strong jet streams arise all the time near the eastern coast of the USA and often over Canada. Over Europe the jets most often form in the region of the British Isles.

Regions with the highest occurrence of jet streams coincide with regions of large horizontal temperature gradients, i.e., during the winter at the junction of the cold continents of Asia, North America, and also Greenland on the one hand, and the warm oceans, on the other hand. A high frequency of tropical jet streams is characteristic of North Africa and of South Asia.

Low frequency of tropical jet streams occurs in regions with a more or less uniform underlying surface. These are the oceans south of 30°–40° NL and north of 30°–40° SL, the northern parts of Asia and America, with the adjacent Arctic regions, and the southern polar region — Central Antarctica.

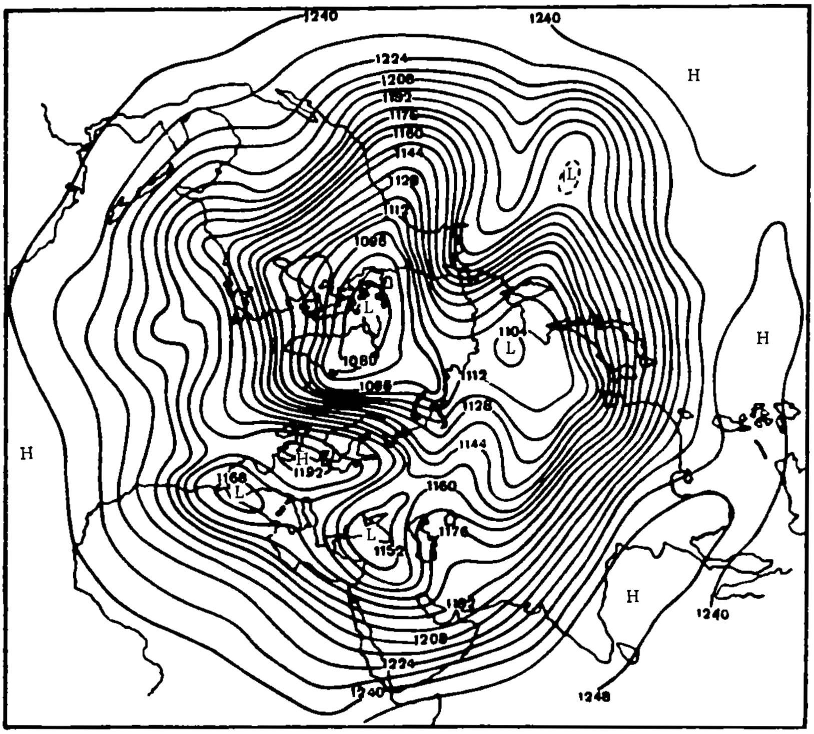

FIGURE 69. Absolute barometric-topography map of the 200-mb surface, 5 January, 1956

Jet streams are usually represented in the horizontal and vertical planes. The wind velocities are then represented by isotachs, i.e., by equal-velocity lines.

Figures 69 and 70 show the maps of the absolute barometric-topography of the 200-mb surface for two different days. The first map is for the middle of the winter, the second — the middle of the summer. The map of the barometric topography of the 200-mb surface (height about 12 km) represents the distribution of the maximum wind velocities in the upper troposphere and in the lower stratosphere. It can easily be seen that a zone of dense isohypses clearly stands out on the background of the sparse isohypses encircling the whole of the northern hemisphere. The highest wind velocities — jet streams — are observed in these zones. At places where the jets fuse the wind velocities increase, and at places where the jets branch off, the wind weakens.

On the evening of 5 January, 1956 (Figure 69) strong jet streams appeared between Iceland and Scandinavia where the southwesterly and northwesterly currents merge. Such strong jets can easily be noticed over south and southeast Asia, Alaska etc. The concentration of the isolines, i.e., high wind velocities, can almost constantly be observed during the winter months to the south of 40° NL (subtropical jets), whereas at moderate and high latitudes, particularly over the USSR, the jet streams become weaker, disintegrate, and again arise with the appearance and development of cyclones and anticyclones.

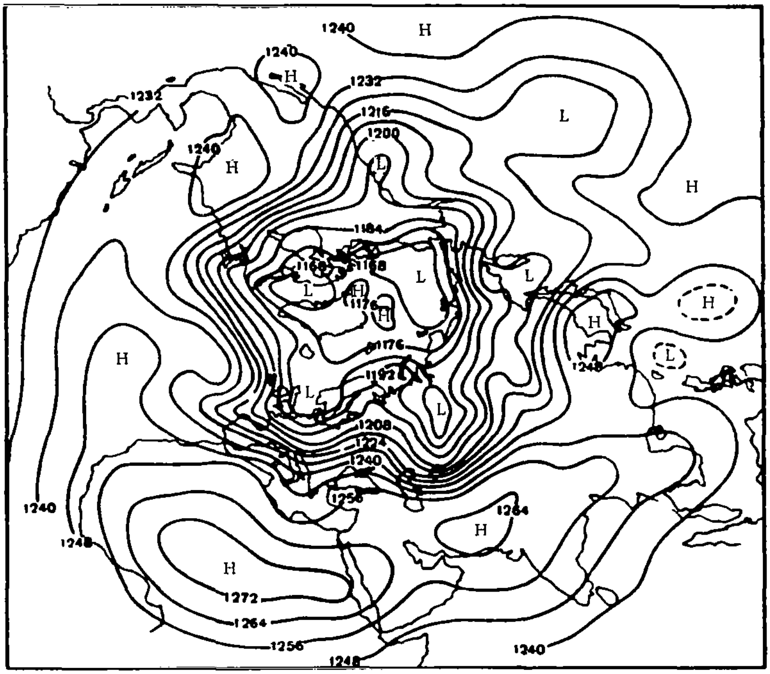

FIGURE 70. Absolute barometric-topography map of the 200-mb surface, 31 July, 1956

During the summer jet streams are encountered very rarely to the south of 40° NL, but are observed more often at moderate and high latitudes. A typical distribution of jets in the northern hemisphere during the summer is shown in Figure 70. It can be seen that the zone of isohypse concentration and of strong winds at the 200-mb isobaric surface passed through moderate latitudes of the northern hemisphere and that the winds were weak over the low latitudes and the Arctic. However, on individual days jet streams may be intensive also at high latitudes.

The spatial structure of jet streams is also shown in a vertical plane perpendicular to the direction of the current. These are usually vertical cross sections of the atmosphere with isotherms and isotachs and of the tropopause. Figures 71 and 72 give two typical examples of vertical cross sections of jet streams for the winter and summer. In these cross sections subtropical and extratropical jets are shown. The letters at the centers of the jet streams indicate the main directions of the air currents.

In Figure 71 two westerly jet streams are shown with their axes lying at the levels of 10 and 12 km up to a height of approximately 25 km between the Equator and the North Pole. The mean maximum wind velocities on the axis of the subtropical jet (to the left), reaching 180 km/hr, were observed over Iraq. The second jet (to the right), was over Moscow at a level of about 9 km. There the mean maximum wind velocities were 100 km/hr, yet at ground level the mean wind velocities did not exceed 10-20 km/hr. During the summer (29 August, 1957) the subtropical jet was situated over Transcaucasia, and the extratropical jet — over Moscow. In the first jet the maximum velocity reached 140 km/hr, and in the second, 120 km/hr.

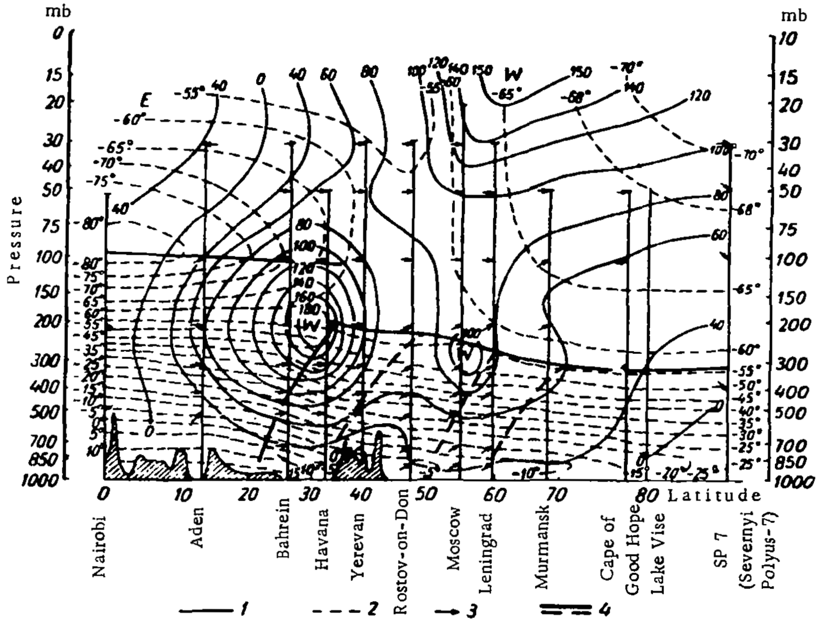

FIGURE 71. Monthly mean vertical cross section of the atmosphere between the Equator and the North Pole, January, 1957–1959

1 – isotachs, km/hr; 2 – isotherms; 3 – direction of the air currents; 4 – tropopause and zones of tropospheric fronts.

In spite of the typical cross sections represented here the positions of jet streams may be different for individual periods.

In view of the considerable difference between the horizontal and the vertical dimensions, the usual stretched out form of the jet is not displayed on the cross sections shown here. However, if we take into account the fact that, for example, in the system of the southern jet in Figure 71 the distance between the low and high positions of the 100 km/hr isotach is approximately 10 km vertically and over 2000 km horizontally, it becomes clear that the jet has a form of a rather stretched out ellipse. The ratios between the vertical and the horizontal extensions in other jet streams are similar.

The characteristic structural features of high-level frontal zones and of jet streams do not undergo noticeable seasonal variations. The seasonal differences are expressed mainly in the intensity and in the latitudinal position of the southern (subtropical) jets.

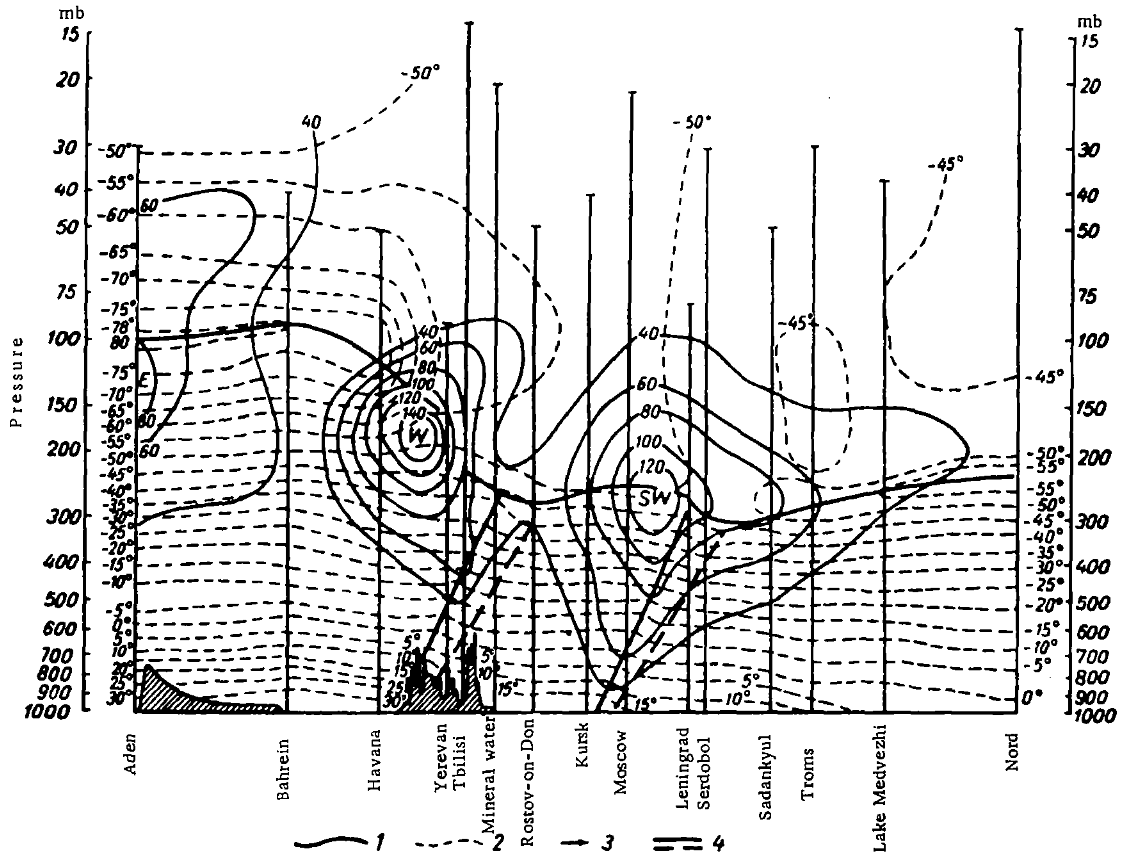

FIGURE 72. Vertical cross section of the atmosphere between the Equator and the North Pole, 29 August, 1957

For notation see Figure 71.

Due to the large temperature contrasts between low and high latitudes, the wind velocities in a jet during the cold time of the year are higher than during the summer, the maximum velocities being observed at lower levels. During the summer the wind velocities are lower, and the maximum velocities are observed at higher levels than those of the winter. Subtropical jet streams have interseasonal displacements along the meridians. This can also be seen in Figures 71 and 72.

In addition, in the system of a subtropical jet stream the tropopause is always torn and the jet axis is situated between the tropical and extratropical (polar) tropopauses. In the zone of an extratropical jet stream, the tropopause is as a rule inclined, it is rarely torn, and the jet axis is most often situated under the tropopause. Therefore at low latitudes the zone of maximum wind velocities is usually higher than at middle and high latitudes. The tearing up and inclination of the tropopause are also displayed on the above given cross sections of the atmosphere.

Some data on the vertical and horizontal extensions of tropospheric jet streams as well as on the mean and maximum velocities in their systems can be found in Tables 27 and 28.

TABLE 27

Vertical and horizontal extensions of extratropical and subtropical jet streams over Europe and Asia, within the limits of the 100 km/hr isotach

Jet streams	Extension, km									
	vertical					horizontal				
	less than 2	3–7	8–12	13–16	over 17	less than 300	301–1000	1001–2000	2001–3000	over 3001
Extratropical	1	25	30	19	10	1	35	40	12	—
Subtropical	—	7	31	31	12	—	5	46	26	6

It follows from Table 27 that subtropical jet streams are relatively powerful and thick. Subtropical jets of large vertical and horizontal extensions (within the limits of wind velocities of over 100 km/hr) are encountered more often than such extratropical jets.

TABLE 28

Mean maximum velocities on the axis of extratropical and subtropical jet streams

Wind velocity, km/hr	Number of cases		
	extratropical streams	subtropical streams	total
100–150	33	8	41
151–200	41	37	78
201–250	13	23	36
251–300	1	12	13
Over 300	2	1	3
Total	90	81	171

Subtropical jets over 2000 km wide and over 12 km high are encountered considerably more often than extratropical jets. However, in individual cases, extratropical jets are powerful and thick, the wind velocities at the center of the jet sometimes attaining 400 km/hr and more.

The mean maximum velocities in systems of extratropical jet streams are most often 150–250 km/hr and in subtropical jets 200–300 km/hr. In other words, with respect to the maximum velocities at the center, subtropical jets are on the average more intense than extratropical jets (Table 28).

Jet streams and cyclones (anticyclones)

It has been established that jet streams are closely related to developing cyclones. In fact, cyclones appear under high-frontal zones where the wind velocities usually attain the highest values, i.e., under jet streams. However, the formation and development of cyclones, as well as of jet streams, are basically determined by the same processes, giving rise to a variation in the horizontal temperature gradients and the structure of air currents.

When we discussed atmospheric fronts, we saw that they form mainly upon a horizontal approach of air masses with different temperature properties.

Analysis of high-level weather maps showed that tropospheric jet streams were closely related to atmospheric fronts, well displayed in the temperature field in the whole troposphere. On the other hand, in the system of a rising cyclone sharply defined fronts always exist. The jet streams are situated in a definite position with respect to them. If we represent the frontline at the surface of the Earth and the jet axis in a horizontal projection, the latter will always be behind the cold front and ahead of the warm front.

Figure 73 shows the surface barometric field in various development stages of a cyclone with the lines of the cold and warm fronts at the surface of the Earth. Each scheme shows the axis of the jet stream, situated near the tropopause at heights of 8–10 km.

As can be seen, in the initial stage of the cyclone (Figure 73 a) the jet axis is situated behind the cold front and ahead of the warm front and somewhat to the left of the surface center of the cyclonic disturbance (in the direction of the stream). In other words, in the first stage of the development of the cyclone the high-level frontal zone with large temperature contrasts and wind velocities is situated directly over the surface cyclone.

Due to the intensive horizontal transport of cold air in the tail section of the cyclone and its additional adiabatic cooling, the high-level frontal zone and the jet stream are displaced to the right of the surface center. This process is associated with a deepening of the cyclone (Figure 73 b).

The cyclone in the troposphere is filled up with cold air as it gets deeper. As a result, the frontal zone together with the fronts and the jet stream pass to the periphery of the filling-up cyclone. In the last stage of development the cyclone loses connection with the jet stream. In all the stages of development of the cyclone, the axis of the jet stream is almost at the same distance from the front lines at the surface of the Earth.

As the systems of anticyclones intensify, the jet stream becomes weaker and erodes.

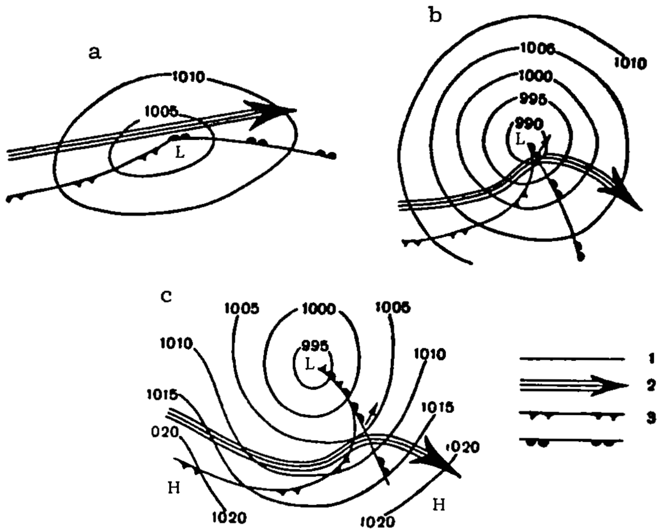

FIGURE 73. Jet streams in the system of a cyclone at various stages of its development

a – appearance; b – maximum deepening; c – filling-up; 1 – isobars; 2 – jet stream axis; 3 – cold front; 4 – warm front.

Winds in the stratosphere and in the mesosphere

It has already been said that above the tropopause the wind usually becomes weaker with increasing height. However, cases were observed in which the wind velocity increased with height in the stratosphere. Such cases were usually observed during the winter at high latitudes, the wind direction being as a rule westerly, and during the summer at low latitudes, with an easterly wind direction.

At the beginning of the 1950's new investigations into the wind regime in the upper stratosphere were carried out. The observed strong winds in the stratosphere were called stratospheric jet streams.

Later it was found that at heights of 25–30 km in the latitudinal zone of 50°–75° of the northern and southern hemispheres small cells of maximum wind velocities appear under certain circulation conditions. It was also found that above 25–30 km in this latitudinal zone the wind velocity continues to increase.

Winds in the stratosphere display a clear seasonal character. During the winter, when the air is highly cooled during the Polar Night, the temperature drops with increasing height down to -70°, -80°C at high latitudes in the stratosphere. In accordance with the direction of the horizontal temperature gradient (from middle to high latitudes) in the above-mentioned latitudinal zone, westerly winds intensify with increasing height and already near the 30-km level reach velocities of 160–200 km/hr.

Figures 74 and 75 give the maps of the barometric topography of the 10-mb surface (according to the author), corresponding to the pressure field at the level of about 30–31 km. It can easily be seen in Figure 74 that as compared with the similar maps for the 300- and 200-mb surfaces

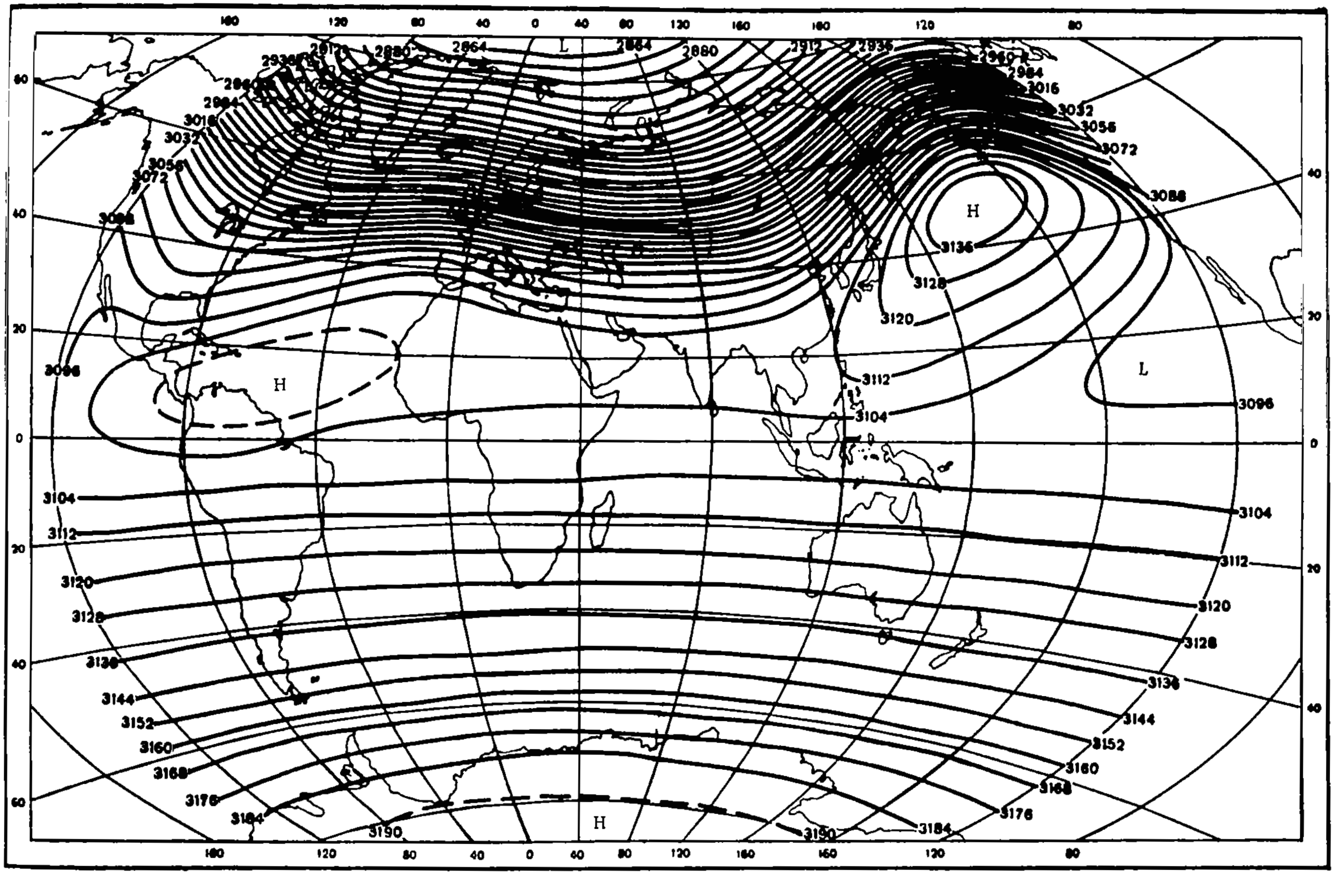

FIGURE 74. Absolute barometric-topography map of the 10-mb surface, January

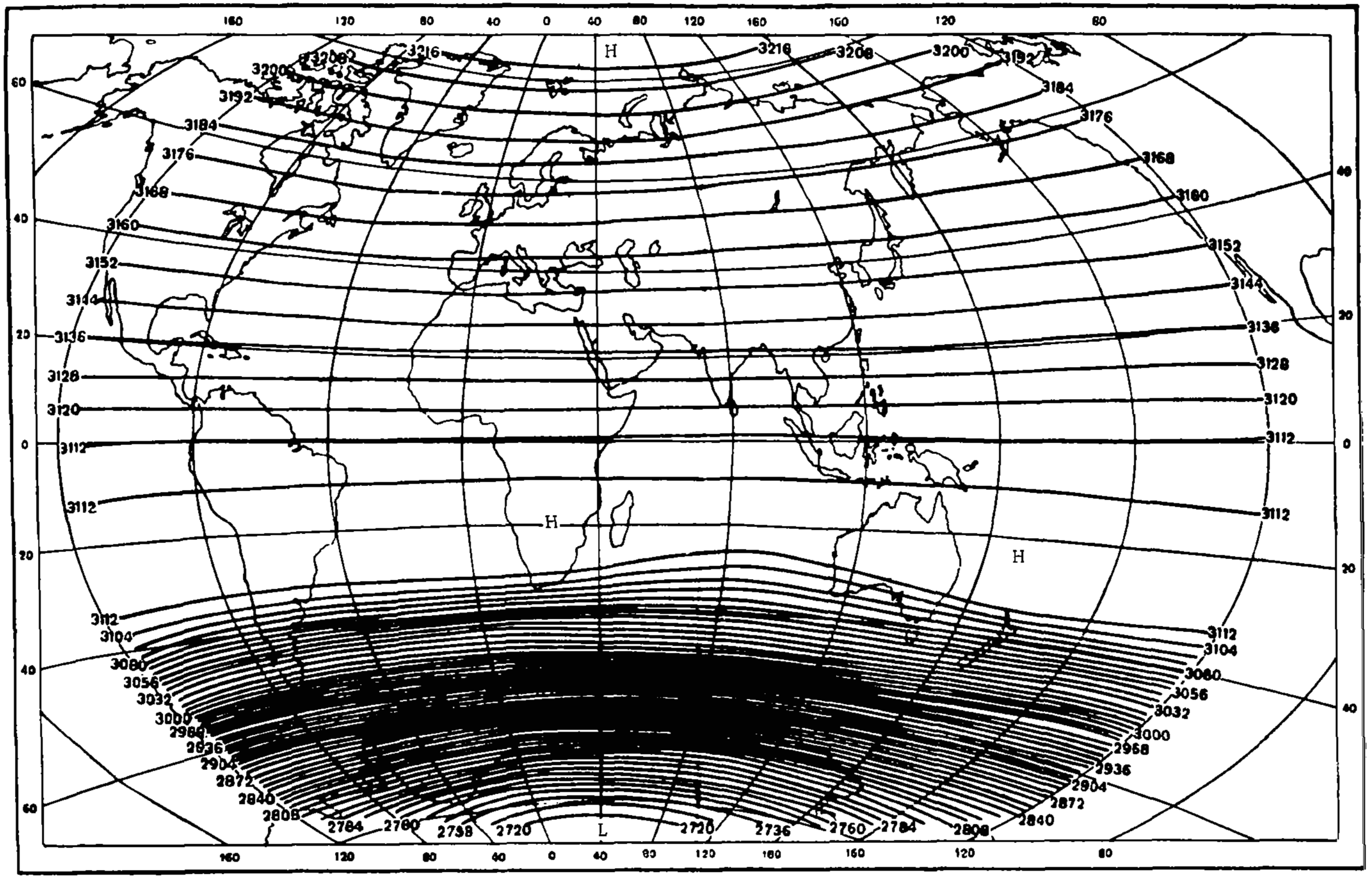

FIGURE 75. Absolute barometric-topography map of the 10-mb surface, July

(see Figures 33 and 69) the structure of the pressure field in the northern hemisphere changed sharply. The high isohypse density between middle and high latitudes indicates high mean wind velocities. To the south of 40° NL on the other hand, the isohypse concentration disappears and weak winds are observed.

At the same time relatively weak easterly winds are observed in the southern hemisphere at the 30–31-km level. This is due to the fact that during January is the Polar Day in the Antarctic, i.e., the air in the stratosphere heats up. We recall that at this level the mean air temperature during the summer months reaches -30°, -35°C and the temperature gradient is directed from high to low latitudes. Westerly winds, characteristic of the troposphere, become weaker above the tropopause and at the level of about 20 km change into easterly winds.

The inverse picture is observed during July. In the northern hemisphere it is the Polar Day, which gives rise to high temperatures in the stratosphere of the Arctic and to a temperature gradient directed from high to middle latitudes. Westerly winds, characteristic of the troposphere, therefore become weaker with increasing height and at the level of about 20 km change into easterly. The low wind velocities in the northern hemisphere are indicated by the low isohypse density there (Figure 75).

In the southern hemisphere on the periphery of Antarctica (similarly to the winter in the northern hemisphere) on the other hand, considerable concentration of the isohypses is observed, indicating high westerly wind velocities in the stratosphere.

For a clearer idea of the characteristic outlines of the horizontal circulation of the atmosphere between the surface of the Earth and the upper stratosphere we give two schemes of the prevailing horizontal circulation (according to the author). The schemes are drawn in accordance with data of daily aerological observations from over 40 stations situated along various meridians (the east of the Atlantic and the west of the Pacific oceans) between the North and South Poles. For this purpose use was made of data for December–February and June–August, averaged over 2–3 years.

The winter circulation in the northern hemisphere is shown in Figure 76. As in the above-given cross sections of the atmosphere, we give the temperature and wind fields, as well as the jet streams. It can be easily seen that on 10 January, 1959 in the northern hemisphere the maximum velocities at the center of the jet along the western boundary of Europe exceeded 110–120 km/hr, and on 11 January on the east of Asia they reached 300 km/hr. In addition, in the stratosphere of the northern hemisphere between middle and high latitudes the velocities of the westerly wind increased with height and at the upper level exceeded 140–160 km/hr.

In the southern hemisphere on these days the mean maximum velocities in the system of the tropospheric jet streams did not exceed 100–140 km/hr and in the stratosphere the winds were weak everywhere. The westerly winds changed into easterly winds near the 50-mb surface (20 km).

During July the situation changes sharply. In the northern hemisphere during the summer the maximum wind velocities in the system of the tropospheric jet are considerably lower than during winter. In addition, the winds in the stratosphere are weak, and above the 50-mb level easterly winds prevail. In the southern hemisphere on the other hand, the mean wind velocities increase with height in the stratosphere, and in the upper stratosphere the (westerly) wind velocities exceed 170–180 km/hr (Figure 77).

There are considerably less data on the wind at heights greater than 30 km. Its direction and speed are determined mainly by sound metering and by launching rockets with smoke charges. Available data on the wind distribution between 30 and 60–70 km agree with the temperature field. On the whole, they characterize the prevailing air transport of various seasons. This cannot be said about the layer above 70 km. There the wind variations over a period of 24 hours are large. There are many contradictions among existing data on the wind direction and speed above the 70-km level which cannot be explained. Therefore, even the mean seasonal characteristics of the wind in the upper mesosphere, and all the more so in the thermosphere, are approximate. An idea about the wind in the mesosphere can be obtained from Figure 78.

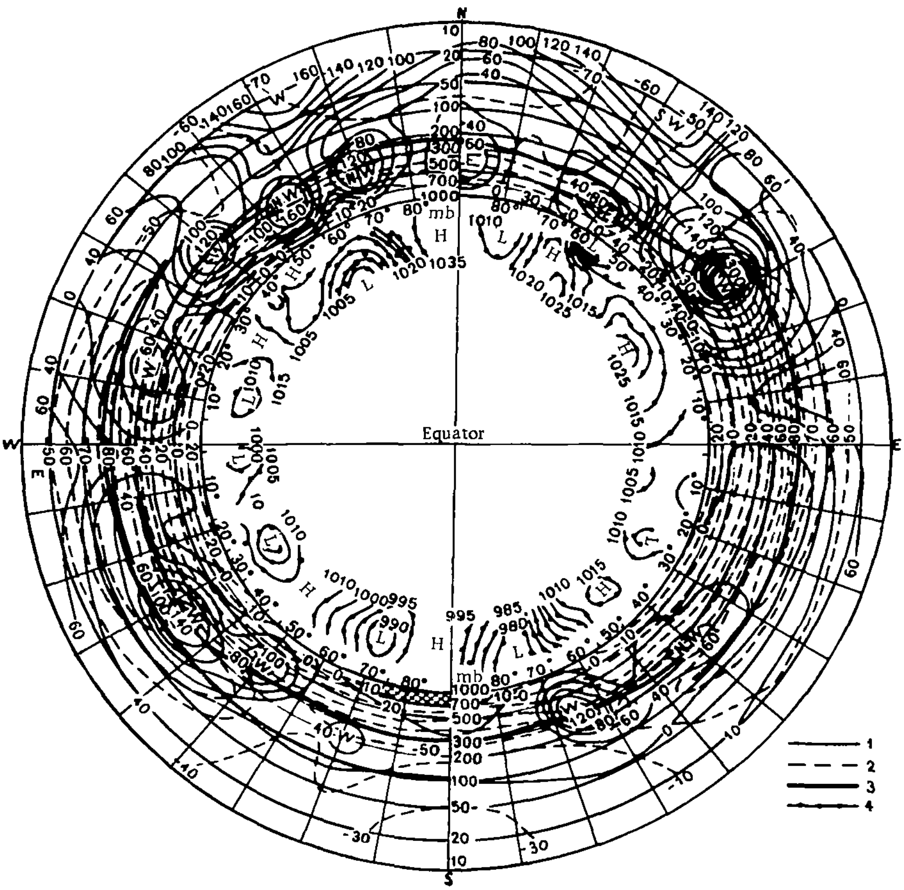

FIGURE 76. Vertical cross section of the atmosphere up to the level of the 10-mb (29–31 km) surface between the North and South Poles along the eastern part of the Atlantic and the western part of the Pacific oceans, 10–11 January, 1959

1 – isotachs, km/hr; 2 – isotherms; 3 – tropopause; 4 – surface isobars along the line of cross section.

Judging from the seasonal mean vertical cross section (Figure 78), westerly winds which reach the maximum velocities (over 350 km/hr) near the 60-km level prevail during winter in the mesosphere. Higher up the velocities decrease, and the winds become easterly in the lower thermosphere.

During the summer, on the other hand, easterly winds (in Figure 78 the isotachs are given with a minus sign) with maximum speeds of about

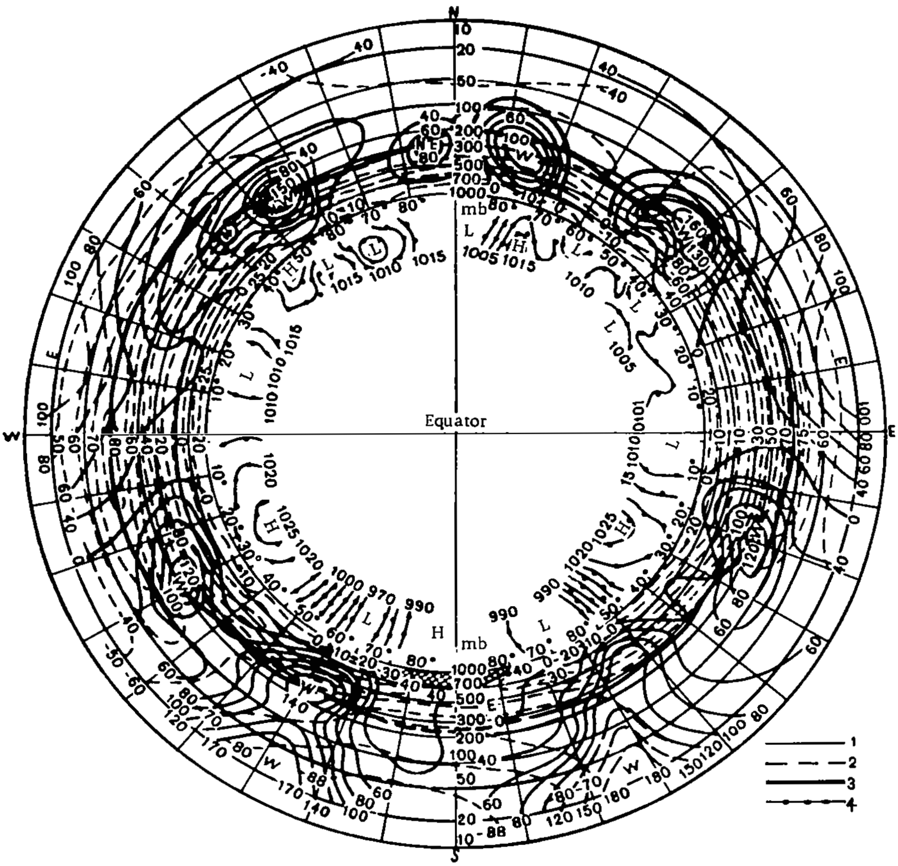

FIGURE 77. Vertical cross section of the atmosphere up to the level of the 10-mb (29–31 km) surface between the North and South Poles along the eastern part of the Atlantic and the western part of the Pacific oceans, 3–5 July, 1959

For the notation see Figure 76.

250 km/hr in the 50–60 km layer prevail in the mesosphere. Higher up the winds become weaker and change into westerly at the boundary of the thermosphere.

If we examine the structure of the wind field on a planetary scale, then in analogy with the wind in the troposphere, the wind system in the mesosphere can be called mesospheric jet streams. However, this problem as well as other problems arising in connection with the peculiarities of the wind regime at great heights, requires a detailed study on the basis of numerous instrumental observations of the wind and temperature.

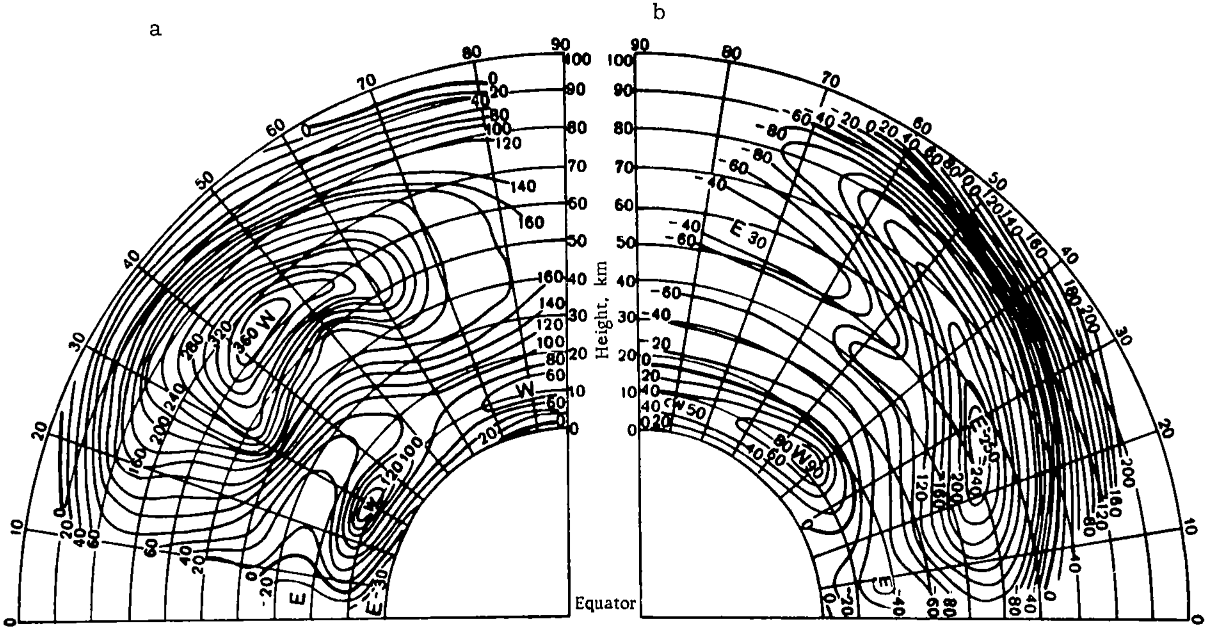

FIGURE 78. Meridional cross section of the atmospheric circulation during the winter (a) and summer (b) up to a height of 100 km

THE WEATHER AND ITS PREDICTION

Weather is the term applied to the state of the atmosphere brought about by physical processes taking place there when the air interacts with the surface of the Earth. The elements which characterize the weather are: the air temperature and humidity, atmospheric pressure, cloudiness, precipitation, wind, snowstorms, fogs, squalls, etc.

The weather exerts a tremendous influence on the economic and cultural activity of man. It is difficult to name any branch of the national economy which does not depend on the weather caprices and does not need its prediction. Weather forecasting is needed for sowing and harvesting of crops, for the provision of trouble-free sea and river transport, telephone and telegraph communications, etc. Information about weather changes is used by everyone in everyday life.

It is most important to predict those weather phenomena which are dangers to the national economy. Knowing the probability of their appearance, measures guaranteeing, if not complete prevention of the dangerous phenomena, at least a reduction in their harmful effect can be taken. For example, receiving a warning of night frosts during spring, flowering fruit trees and young seedlings can be warmed. Information on snowfalls or snowstorms gives time to quickly clear up the snow deposits from railroad lines and highways. Knowing the weather along the flight route, the pilot can choose the best height for the flight, thus avoiding thunderclouds, and icing zones.

The science of studying atmospheric processes with the purpose of predicting the weather is very interesting and at the same time very complicated. Daily the multimillion radio audience of our country listens attentively to the communications of the Central Forecasting Institute and of the local Weather Bureaus about the expected weather for the coming days. Many listeners have no idea that the short weather forecast is produced by the hard work of a group of specialists headed by a physicist forecaster (synoptic meteorologist) and numerous assistants, working not only side by side with them, but at all the corners of the USSR and of the world.

The weather at middle latitudes is highly variable. Everyone knows from personal experience that clear, sunny weather is very often rapidly replaced by cloudy or dull weather with rain; warm weather is replaced by cold weather, a calm – by a strong wind. The downpours and thunderstorms, snowstorms and fogs, dry winds and dust storms, and other weather phenomena harmful to the national economy occur frequently. At middle and high latitudes the weather variability is to some extent characteristic of all the times of the year. However, extensive regions exist where the weather varies little from day to day. For example, in the tropics the character of the weather varies little during a season, and the variation of the weather from season to season takes place in a definite order.

Organization of a storm warning service

Systematic observations of the state of the atmosphere, or, as they are usually called, meteorological observations, began 200–300 years ago, after the invention of the thermometer and barometer made it possible to measure the temperature and pressure of the air.

Already in the 18th century much attention was paid to the study of atmospheric processes. For example, M.V. Lomonosov not only invented various meteorological instruments, but personally conducted observations on the weather, studied atmospheric electricity, thunderstorms, etc. He suggested creating a network of meteorological stations to predict the weather on the basis of a simultaneous survey of the atmospheric parameters over a large territory.

This idea of carrying out simultaneous observations of the state of the atmosphere to forecast the weather for practical problems of sea navigation and agriculture became widespread. The speedy organization of a network of meteorological stations, and the accumulation of observational data made it possible to begin a regular compilation of synoptic weather maps. Such maps were published for the first time in the middle of the 19th century in Europe and in North America. From these maps it was noticed that warm and cold weather depend not only on the direct inflow of solar energy, but also on the motion of air currents. With the aid of these maps extensive regions of low and high pressure, and cyclones and anticyclones with their vortex-like character of air currents were discovered. The relation between the wind, the pressure field and other phenomena was established. However, at that time the weather maps could not be used for getting information on the current weather, since they took a long time to prepare. The rapid collecting of information on current weather became possible with the invention of the telegraph.

One episode from the Crimean War between Russia and the coalition of England, France, Turkey, and Sardinia (1853–1856) served to accelerate the preparation of current-weather maps to be used for storm warnings. On 14 November, 1854 a very strong storm raged in the Black Sea which caused the destruction of the Anglo-French fleet standing at Balaklava. It was established that such a violent storm had appeared the day before in the region of the Mediterranean Sea. Interested in the course of the events, the French astronomer U. Leverrier observed by means of weather maps that the storms which broke out on the Mediterranean Sea and at Balaklava had the same origin: a huge cyclone over the Mediterranean Sea which later moved to the Black Sea and caused the catastrophe. This fact gave the idea that approaching storms can be detected before they arrive, and consequently, that a warning is possible. With the establishment of the possibility of storm warnings, the operational collection of meteorological observational data and the preparation of synoptic weather maps were quickly organized.

A special scientific meteorological center – Central Physical Observatory which headed the organization of a network of meteorological stations and already in 1856 prepared synoptic weather maps, was in 1849 set up for the first time in Russia. But a systematic preparation of these maps and a regular publication of weather data began only in 1872, when the first bulletin describing weather at 20–30 places in Europe was issued.

The basic work of the weather service at that time consisted in issuing storm warnings to the vessels of the Baltic Fleet. Weather forecasting in the modern sense could not even be considered. This is obvious, since the forecasters of that time did not have the experience of forecasting work with weather maps, did not know many of the laws of the development of atmospheric processes, and had at their disposal very scanty data of meteorological observations, carried out only at the surface of the Earth for weather analysis and forecasting. Among the Russian meteorologists, a special place is occupied by P.I. Brounov (1852–1927). To him belong a series of radical developments, which even today find practical application in forecasting work.

Weather forecasting in the USSR

The meteorological network in Russia was almost completely destroyed during the first world war and the Civil War. The young Soviet State in the first days of its existence took effective measures for the restoration and wide development of the network of meteorological stations. On 21 June, 1921 V.I. Lenin signed the decree of the Council of People's Commissars on the organization of a meteorological service in the RSFSR. The result of this decree was a rapid increase in the number of meteorological stations and observatories, which were built in various parts of the country. New scientific-research institutions were established to find solutions to meteorological and their closely related hydrological problems.

By the decision of the Soviet Government the Hydrometerological Committee of the USSR was created in 1929 to unify and improve the organization of all meteorological and hydrological research. Later on this committee was reorganized, and its functions are now carried out by the Central Board of the Hydrometeorological Service of the USSR Council of Ministers.

The weather service in the USSR then grew so much that it began to provide weather forecasts for all branches of the national economy. As the national economy of the country, and particularly aviation developed, the requirements from the weather forecasters increased. These requirements forced the meteorologists to pass from general, qualitative formulations of the character of the expected weather of the next few days, to more concrete indications of the quantitative characteristics of the expected elements, namely: the temperature, wind speed and direction, visibility, and other weather elements and phenomena. As a result of the considerably expanded network of meteorological stations, of the appearance of new means and types of observations, as well as of the great scientific advances, it became possible to prepare weather forecasts in much more detail. The analysis and forecasting of the development of atmospheric processes on the whole and of the weather in particular improved.

In operational forecasting much depends on the data of observations carried out at meteorological stations in all parts of the world arriving on time. Therefore, parallel with the considerable expansion of the network of meteorological stations, the rapid arrival of telegrams on the weather observed simultaneously at many points had to be arranged. The development of all means of communication, and primarily of radio communications, guaranteed a steady and timely arrival of these data not only from populated places, where most of the meteorological stations are situated, but also from far and sparsely-populated regions, in particular from the Arctic and from the Antarctic, from taiga stations, mountain passes and ships on the high seas. Radio communication also guarantees the arrival of meteorological observations from foreign countries in the northern and the southern hemispheres.

In addition to meteorological observations carried out at the surface of the Earth, wide use is now made of aerological observations characterizing the state of the atmosphere at various heights, beginning from a few hundreds of meters to an altitude of 25–35 km. These include observations of the wind speed and direction high up, of the distribution of the air pressure, temperature, and humidity, observations of the thickness of cloud layers and so on. As was already said in the chapter "The Earth's Air Envelope," aerological observations are carried out by means of pilot-balloons, radiosondes, and special airplanes lifting meteorographs which record the meteorological elements. All these data of surface and aerological observations are included in the arsenal of the modern forecasting meteorologist and investigator.

Only twenty years ago the sounding of the atmosphere had an episodic character. The first radiosonde was sent from Pavlovsk (near Leningrad) on 30 January, 1930. Today, the regularly working network of radio sounding stations covers almost the entire northern hemisphere. They exist even on the oceans and in hardly accessible regions of the Arctic and Antarctic.

Such observations are also carried out in the southern hemisphere, but on a more limited scale. Data of aerological observations, as well as of ground observations, are transmitted by modern means of communication to the centers of the weather service and are used for weather forecasting.

Aerological observations contributed much to our knowledge of the structure of the lower layers of the atmosphere as well as to the development of methods for weather prediction. In order to forecast the weather many features of the state of the atmosphere must be known, as well as the variation regularities of the processes. Before systematic aerological observations existed, the forecasters had only indirect data on the state of the atmosphere higher up, obtained from observations carried out at the surface of the Earth. It was clear that surface data alone were insufficient to determine the processes developing throughout the atmosphere, in the same way as it is impossible to know the features of the surface of the ocean from its bed.

Aerological observations clarified many complex structural features of the atmosphere and established the character and reasons for the variation of the processes taking place there. They also enabled us to pass from guesswork to actually knowing the conditions of weather formation and to create scientific methods for weather prediction, which is the problem of synoptic meteorology.

Synoptic meteorology uses the laws of physics and hydrodynamics. But the accuracy to within which a given science can predict a phenomenon depends on the degree of perfection which it reaches in its development. Astronomy, for example, reaches a high degree of perfection, but this was not always the case. At the beginning of the development of celestial mechanics the accuracy with which the motion of planets and comets was calculated was considerably lower than now.

Weather forecasting is a much more complicated problem than calculating the motion of planets; it depends on a large number of factors that vary greatly with respect to time and space. Thus, air "carrying a certain type of weather," interacts continuously with the surface of the Earth, or, as it is usually called, the underlying surface. In this interaction its properties vary continuously due to the nonuniform heating and cooling, to the increase and decrease in the moisture content, to the variation in the horizontal and vertical motion, etc.

Modern synoptic meteorology has not yet coped with many of the difficulties arising in weather forecasting. However, further development of this science will lead to successes similar to those attained by astronomy. But first many difficulties will have to be overcome. In contrast to astronomy, synoptic meteorology deals with the equations of the mechanics of a continuous medium. It is difficult to obtain a general solution of the equations of thermo- and hydrodynamics, and we must be satisfied with the solution of particular problems and use qualitative results in cases where the hydrodynamic theory does not apply. Progress is very difficult, but every means is being used to achieve it.

The forecasting meteorologist cannot yet take quantitatively into account all the factors influencing the formation of the future weather nor can he, on the basis of the observation data, calculate the weather elements by means of equations representing the laws of the atmosphere.

The weather and its variations

The weather at any point on the terrestrial globe undergoes variations. Some of these variations are periodic, e.g., those which depend directly on the radiational heat exchange over a 24-hour period or a season. Periodic weather variations are displayed in the daily and in the annual march of the air temperature, humidity, pressure, the wind, and other elements. The largest daily variations of the temperature and relative humidity are observed over land in clear and windless weather. In sunny weather the air is heated by day from the surface of the Earth, and on clear nights, owing to the long-wave radiation, it cools. When the air is heated the relative humidity drops sharply, and rises at night as a result of the cooling. If the moisture content and the cooling of the air are sufficient, condensation of the water vapor begins, forming dew and fog. Over sea the daily temperature variation is very small, since the heating and cooling of the water surface is considerably slower than the land.

The daily variations in temperature, humidity, wind, and other meteorological elements are influenced by the topography and the plant cover. For example, in hollows or over dry soil the air heats-up or cools-down much more rapidly than on slopes or over a humid soil. The daily temperature amplitudes over a soil without plants and over a soil covered by plants are different. The daily temperature and humidity oscillations decrease with increasing height and usually damp out at a height of 1–2 km. But even at a height of over 10–12 km daily temperature oscillations, determined by the radiational regime at these heights, are also observed. The temperature variations from season to season also extend throughout the whole thickness of the troposphere and the lower layers of the stratosphere.

Nonperiodic weather variations are most clearly displayed in extratropical latitudes, where the circulation of the atmosphere is most intensive. They are connected with the variation of the physical properties of the air which in turn depend on the character of the underlying surface over which the air moves and on the character of the atmospheric circulation, the origin of which is the nonuniform distribution of the solar energy over the terrestrial globe. Thus, for example, dry cold air when moving from a cold continent to a warm ocean is heated and moistened; if it stays long over the ocean, convective cloudiness is formed and torrential precipitation falls. When moistened and heated air over the warm ocean moves to the cold continent the temperature of its lower layers drops, the state of saturation sets in, and fog or low-level strati cloudiness is formed. During summer air masses moving over a heated continent are heated and removed from the state of saturation; this leads to clear dry weather.

A variation in the properties of air also takes place when it moves along a meridian over a uniform underlying surface. Thus, for example, when air masses move from north to south they are heated and moistened. When moving from south to north, on the contrary, air masses are cooled and due to the fall of precipitation their moisture content decreases. The sharpest nonperiodic weather variations are connected with the activity of atmospheric fronts, cyclones, and anticyclones.

Until recently the point of view prevailed that the weather depended mainly on the place of origin of the air mass. This is only partially true, however. Of course it is not immaterial from which geographic regions the air mass originates — from cold Siberia or the warm Atlantic. During winter in the first case the air mass carries with it strong frosts and in the second case it causes warming up until thawing sets in. The properties of moving air vary by the interaction with the underlying surface. But another thing should not be forgotten. Air masses may arrive at different times from the direction of the Atlantic Ocean over the European territory of the USSR having the same temperature and moisture content, yet the weather may be different, depending on the circulation system in which the air masses are involved. Thus, in deepening cyclones, due to the convergence of air currents in the surface layer, to the increase in the horizontal temperature and pressure gradients, and to the ascent and dynamic cooling of large air volumes, atmospheric fronts intensify, multilevel cloudiness develops, and prolonged steady precipitation falls. The same air masses in the system of intensifying anticyclones, where descending motion and spreading, and consequently also dynamic heating of the air are observed, cause the fronts in the surface layer to erode, the cloudiness usually scatters, and clear or slightly cloudy weather begins.

Over various geographic regions in extratropical latitudes, the weather is different depending on the position of the cyclones and anticyclones. Since cyclones and anticyclones usually move from west to east with velocities often exceeding 30–40 km/hr or over 700–1000 km per 24 hours, the weather observed over the central regions of the European territory of the USSR may have reached West Siberia 24 hours later. When the cyclones and anticyclones are almost stable, the weather in that region hardly varies.

During the winter in the USSR anticyclones bring slightly-cloudy or clear weather, with weak winds or frosts, and during summer — dry weather with high temperatures are observed. With cyclones during the winter the skies are usually overcast, with weak precipitation and moderate frosts, while during summer there is comparatively cool weather with downpours and thunderstorms when the moisture content of the air is sufficient.

Strong winds cause an increase in the horizontal pressure gradient. This takes place in systems of cyclones when they rapidly deepen or on the periphery of intensifying anticyclones. Often the same type of weather may be simultaneously observed over quite a large territory (with a radius of up to 1000 km and more). But it can often happen that at a very short distance (of a few tens of kilometers) the weather may be very different. Over one part of a small territory overcast weather with precipitation, and alongside cloudless weather can be observed. The temperature difference in this case may reach 10°C and more.

The Atlantic Ocean has a large influence on the weather of Europe and Asia. This is particularly so during the winter, when air masses heated over the ocean penetrate to the interior of the continent. Considerable temperature rises to the north of the European territory of the USSR during winter, reaching thawing in individual cases, occur during a prolonged western transport of warm air masses. Getting farther into the interior of the Asian Continent, the influence of the Atlantic Ocean becomes weaker, and with the frequent development of anticyclones and clear weather in Siberia and Central Asia during the winter the surface of the Earth and the lower air layers cool down considerably. The strongest frosts in Europe are due to the penetration of air masses from Siberia or from the Arctic. During spring and fall returns of cold are caused by the penetration of cold air masses from the north or from the northeast. When conditions exist for radiational cooling (in a system of an anticyclone) strong frosts are formed in the surface air layer. During the summer in Central Asia the air masses are heated more than in Europe.

In contrast to middle and high latitudes, at low latitudes the weather variations are mainly periodic, determined by the change in seasons. With this change and the displacement of the calm zone, the direction of the trade winds and of the air currents between the continents and the oceans (monsoons) change. Individual discrepancies are mainly caused by processes developing in the extratropical latitudes and by tropical cyclones often arising over the oceans.

The weather service and weather forecasts

Weather forecasts are divided into short-range (1–2 days) and long-range forecasts. The latter in turn are subdivided into short-term (3–10 days) and into long-term (month, season). Another division is into general and specialized forecasts.

General weather forecasts contain information on the temperature, cloudiness, precipitation, wind, thunderstorms, squalls, snowstorms, fogs, spring or fall frosts, storms, etc. Specialized weather forecasts are prepared for various branches of the national economy. For example, railroad and automobile transport need to know in advance of snowfalls and snowstorms blocking roads, or of heavy rains washing away roads. Kolkhozes and sovkhozes require information about expected precipitation, dry periods, dry winds during the winter and fall, frosts, etc. Specialized weather forecasts for aviation include data on the visibility, vertical thickness and lower boundary of clouds, wind at various heights, possibility of icing of aircraft wings along the flight route, and so on.

The expected temperature, speed and direction of the wind (in general forecasts), lower and upper boundaries of the clouds, range of horizontal visibility (in forecasts for aviation) are given quantitatively but in fixed scales. The classification of the expected weather involves various dangerous phenomena such as fogs and snowstorms, dust storms and thunderstorms, dry winds, etc.

The monthly and particularly the seasonal forecasts give only a general characterization of the weather and dates of sharp variations of the air temperatures.

The weather service centers are called to supply the country with weather forecasts, as well as with information on the current weather. Weather service centers exist all over the world. In the Soviet Union there are weather bureaus in the capitals of the Soviet Republics and in large regional centers, meteorological and hydrometeorological bureaus in many regions, and aviation meteorological stations at airports. The latter supply information and weather forecasts to aviation. The weather bureau supplies daily weather and hydrological forecasts to industry, agriculture, transport communications, etc. The central scientific research and forecasting institution in the USSR is the Central Forecasting Institute at Moscow.

The weather bureaus and other weather service institutions issue information so that the development of atmospheric processes causing weather variations can be followed all the time. The work is carried out around the clock, and thousands of telegrams containing data on the weather in various countries are received and processed daily.

The modern methods of weather forecasting are based on a simultaneous study of atmospheric processes and the weather variations caused by them. This is done by means of weather maps, which are called "synoptic," i.e., survey maps.

Weather maps

Weather maps are special geographic maps on which, by conventional signs and cyphers, data on the weather observed simultaneously at fixed times at meteorological stations are plotted. The weather service centers receive these data in a coded form. Cyphered telegrams consisting of five-digit groups are transmitted by telegraph, telephone, and radio. These telegrams contain information on the weather at a definite instant at points situated in various geographic regions. These data are plotted by special marks on the map. By drawing isolines, i.e., lines of equal pressure, temperature, etc., on the weather maps, one obtains the fields of these meteorological elements. This makes it possible to determine the regions of high and low pressure, of cold and heat, of dry and humid, stable and unstable air masses, etc.

The weather maps are prepared several times during a 24-hour period and serve, as was already said, as the basic material for weather forecasting. There are basic and auxiliary, surface and high-level weather maps. The basic maps, usually with a scale of 1: 10,000,000, contain information on the weather over an area of 30–40 million km^2. The auxiliary maps contain data on the amount of precipitation, maximum and minimum air temperature, soil temperature, the pressure variation during definite time intervals, and so on.

The scales of the various maps are different. For example, the maps giving a detailed description of the weather in an individual region for which the forecast is made, are of a scale 1:5,000,000 and more. Weather maps of the northern hemisphere, on the other hand, usually have a scale not less than 1:30,000,000 etc.

$T_e T_e$	C_H	
T T	C_M	PPP
w w	(N)	±ppa
VV	$C_L N_h$	W
$T_d T_d$	h	RR

FIGURE 79. Arrangement on a weather map of weather data at an observation point

The number of maps prepared in the weather bureau during 24 hours depends to a certain extent on the term of the forecast. For example, for a daily weather forecast, the basic maps are prepared every 6 hours, the circumferential maps — every 2–3 hours.

For weather forecasting for a long period, surface and high-level weather maps extending over the northern hemisphere are prepared twice a day. The dimension of the territory is determined by the velocity of the atmospheric fronts, and of cyclones and anticyclones carrying some type of weather. During 24 hours they often traverse 1000 km and more.

Data on the weather at a given point are plotted on the surface weather map by the scheme adopted by the World Meteorological Organization. In this scheme (Figure 79) the circle (O) denotes the position of the meteoro-ogical station on the map;

N – the total amount of precipitation;
VV – the horizontal visibility;
ww – the weather at the time of observation or during the last hour before the observation;
W – the weather between the observation times;
PPP – the air pressure;
TT – the air temperature;
N_h – the amount of cloud, whose height is indicated in meters;
C_L – index of low-level clouds;
h – height of low-level clouds;
C_M – index of medium-level clouds;
C_H – index of high-level clouds;
$T_d T_d$ – the dew-point temperature;
± pp – the magnitude of the barometric tendency;
a – index of the barometric tendency;
$T_e T_e$ – the extremal temperature;
RR – the amount of precipitation during the last 12 hours.

The direction (dd) and the speed ($f_m f_m$) of the wind are indicated by an arrow with feathering.

For each symbol a large number of notations exists given by the meteorological code.

From the examples of weather data coding given in Figure 80 it is possible to get an idea of the language of cyphers and signs of the synoptic weather map.

Example a. 8/10 of the sky was covered by cumulonimbus and high-cumulus clouds. The lower boundary of the cumulonimbus clouds was 600 m. At the period of observation a moderate thunderstorm with rain was observed. The amount of precipitation that fell during 12 hours was 10.0 mm. The wind was northwesterly, 14–16 m/sec. The air temperature was 17°C, the temperature of the dew point 12°C. The pressure was 1002.5 mb. During the last 3 hours a slight pressure drop, which afterwards changed into a slight rise was observed. The visibility was over 50 km.

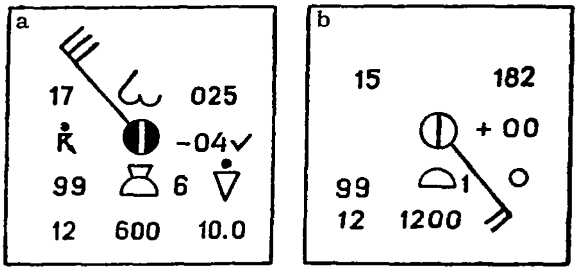

FIGURE 80. Examples on a weather map of "unstable" (a) and "good" (b) weather at an observation point

Example b. 1/10 of the sky was covered with cumulus good-weather clouds the lower boundary of which was 1200 m. The pressure was rising, at the time of the observation it was 1018.2 mb. The air temperature was 15°C, the temperature of the dew point 12°C. The wind was southeasterly, 9–11 m/sec. Between the observation times there was clear weather. The visibility was over 50 km.

As we can see, in the second example the weather had quite a different character.

Thus, by analyzing weather maps with detailed data on the weather in various parts of Europe, Asia, or the northern hemisphere, the meteorologist seems to read the history of the atmosphere. Comparing a series of maps for successive times, he learns about the struggle of warm and cold air masses, resulting in atmospheric fronts, and vortexes of large and small dimensions, carrying with them storms and gales, cold and heat, bad and fine weather, drought and rain.

Figure 81 shows a small section of the surface weather map with data on the weather plotted according to the scheme of Figure 79. The basic map, on which in this way the data of 400–600 stations are plotted, is ready for analysis 3–4 hours after the routine observation has been carried out. To the people at the meteorological stations, at the communication center, and at the weather service centers every minute, every second counts. During a comparatively short time interval the observers carry out a whole set of meteorological and aerological observations, record them in special log books, process and code the obtained information and transmit it by telegrams through the communication centers, to the respective weather bureaus. There they are decoded and plotted on special geographic maps. In particular, the weather bureaus situated in eastern Europe prepare maps which cover the entire territory of Europe, the North Atlantic, western Siberia, Kazakhstan, and Central Asia. In the weather bureaus situated in other geographic regions, other blanks are taken. In a number of weather bureaus even a map of the entire northern hemisphere and a number of other auxiliary weather maps are prepared.

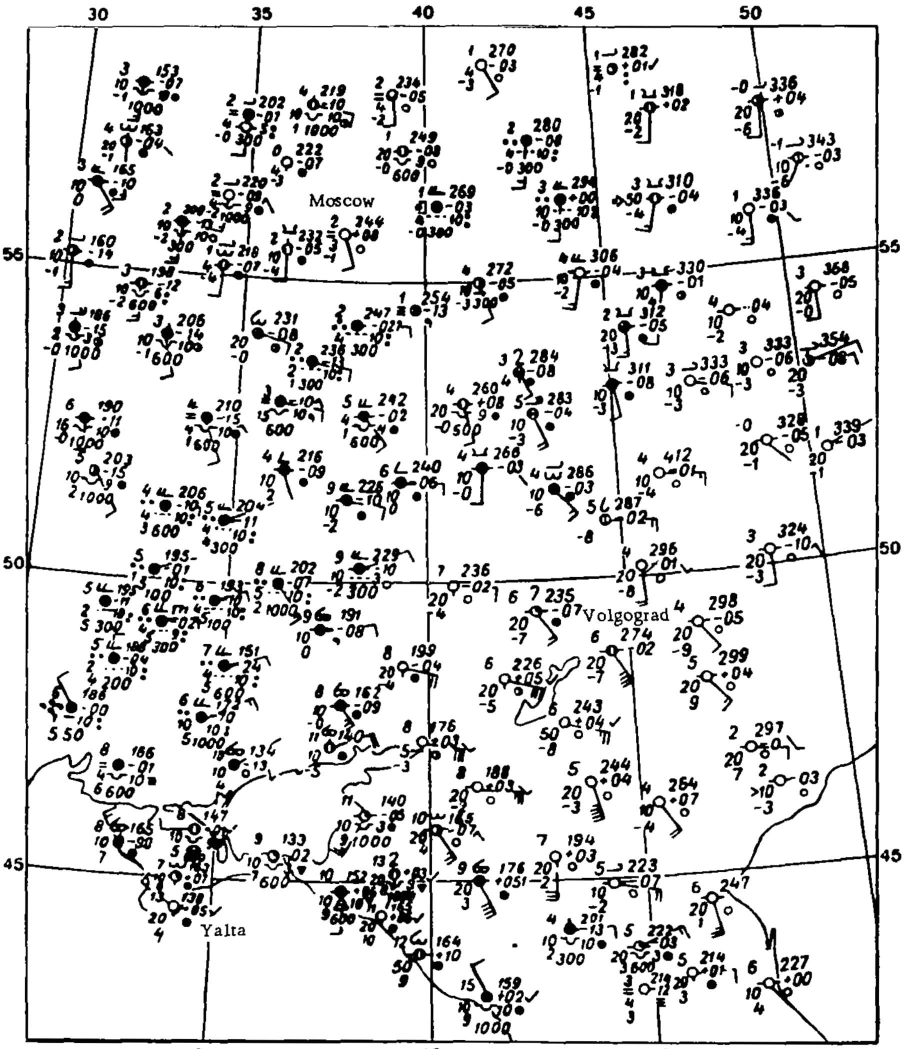

FIGURE 81. A section of the surface weather map for a 21-hr period on 13 April, 1955

Attempts have recently been made to prepare daily maps of the world weather.

Since 1948, to guarantee synchronization of the meteorological and aerological observations, they are carried out simultaneously over the entire territory of the USSR at four basic times, i.e., at 0300, 0900, 1500, and 2100 hours, and at other intermediate periods (Moscow time).

High-level weather maps. High-level weather maps are also prepared two to three times a day. On these maps data of aerological observations carried out at fixed times in all the countries of the world are plotted. High-level maps contain data on the pressure, temperature, and humidity of the air, and on the speed and direction of the wind at various heights.

The high-level maps are used in the first place to represent the pressure field, i.e., the pressure distribution at various levels. The pressure field is represented by two methods, one of which we discussed above. By this method the barometer readings at the meteorological stations (with the corrections for comparability) are reduced to one level. In particular, on the surface maps the air pressure at sea level is plotted. The same can also be done on the high-level maps by plotting pressure data at a fixed height, for example, at the level 1, 2, 3, . . . , 10, . . . , 20 km and so on.

To represent the pressure field at various heights another method is used at all the weather services around the world. This is the so-called topography method of isobaric surfaces, or the barometric topography, by means of which the surface of equal pressure, or of an isobaric surface is represented. We discussed such maps in the chapters on "Motion of the Air" and "The Vertical Temperature Lapse Rate" in the study of the distribution of the pressure and of the air currents in the troposphere, i.e., at a level of about 9 km, and of the temperature in the layer between the 1000- and 300-mb surfaces, characterizing the mean temperature of the lower 9-kilometer layer of the troposphere.

The theoretical basis for preparing high-level weather maps, and in particular for the growing method of barometric-topography maps, was worked out by V. Bjerknes in 1912, long before the appearance of operational methods of atmospheric sounding, and the use of aircraft and radiosondes in investigations of the free atmosphere.

Barometric topography maps are usually drawn for the principal isobaric surfaces. Principal isobaric surfaces are the surfaces of equal pressure, of 1000, 900, 850, 700 mb and so on. The 1000-mb surface is situated near sea level; the 900 mb surface corresponds to a height of about 1 km; 850 mb – about 1.5 km; 700 mb – 3 km; 600 mb – 4 km; 500 mb – 5.0–5.5 km; 400 mb – 7 km; 300 mb – 9 km; 200 mb – 11.5–12.0 km; 100 mb – 15–16 km.

The heights above sea level, calculated from sounding data of each aerological station taking into account the distribution of the air pressure, temperature, and humidity are plotted on the isobaric surface map.

The height of the isobaric surfaces is calculated not in geometric meters but in geopotential meters.*

In order to calculate the geopotential the acceleration due to gravity must be known. The magnitude of the acceleration due to gravity is not constant over the terrestrial globe. At the Equator it is somewhat lower than over the Poles. At 45° latitude the acceleration due to gravity is equal to 9.80616 m/sec^2. The work is measured in extremely small units – ergs.

* The gravity potential or the geopotential is the work which has to be done in order to lift a unit mass from sea level to the given height.

An erg is the work done by a force of 1 dyne in moving 1 cm in the direction of the force. The unit of the geopotential is equal to 100,000 erg/gram, and is called a dynamic meter (dyn. m). A dynamic meter is the energy spent in lifting a mass of 1 gram to a height of 1 m. At 45° latitude this work is approximately equal to 0.98 dyn. m. Hence it follows that 1 m of length is equal to a geopotential of 0.98 dyn. m, or that 1 dyn. m corresponds to a height difference of 1.02 m. To be able to express the variation of the atmospheric pressure with height in terms of the geopotential considerably simplifies the calculations, since in this case only data on the pressure and temperature are required in order to set up the pressure and air-current field.

However, certain inconveniences arose in the practical work of meteorologists due to the fact that a difference of 2% exists between the geometric and dynamic meters. Therefore the World Meteorological Organization decided in 1950 to adopt a new unit of height, more convenient than the old one since it is very close to the geometric meter. The difference between the geometric meter is negligible, not exceeding 3–4 m per 1 km depending on the deviation of the acceleration due to gravity from the value of 9.8 m/sec^2. The new unit is called a geopotential meter.

The absolute geopotential of an isobaric surface depends on the surface pressure and on the mean temperature of the layer. The relative geopotential depends only on the mean temperature of the layer between the isobaric surfaces. Accordingly, absolute and relative barometric-topography maps are used.

The heights of this surface above sea level in geopotential meters or decameters are plotted on absolute topography maps. Connecting points of equal heights, i.e., drawing isohypses (as it is done when drawing the topography of the surface of the Earth), we obtain the topography of the isobaric surfaces. On these maps it is possible to see elevations and depressions, to which anticyclones and cyclones, crests and troughs, and zones of higher and lower isohypse density correspond. Since the isohypses are equivalent to isobars, the absolute topography maps represent the pressure field at the given level. The absolute topography maps contain data on the air temperature and humidity at a given isobaric surface, as well as the wind speed and the direction from pilot-balloon observations and the dew-point temperature.

The layer thicknesses between two isobaric surfaces are represented by relative topography maps. Since the heights depend only on the mean temperature of the layer, the relative topography maps essentially represent the features of the temperature distribution in the air layer under consideration. One can observe regions of cold and heat, small and large horizontal gradients in frontal zones, etc.

In the weather service of the USSR absolute topography maps of the 850, 700, 500, 300, 200, and 100-mb surfaces, and a relative topography map of the 500-mb isobaric surface are prepared. Topography maps are also prepared for special purposes for higher lying surfaces, up to the 30- and 10-mb levels, which correspond to heights of 24 and 30 km.

Barometric-topography maps make it possible to determine the thermobarometric structure of cyclones and anticyclones, and of frontal zones and atmospheric fronts. The main currents in the atmosphere may be found in this way if pilot-balloon data are missing. They play a very important role in the determination of the state of the atmosphere, and make it possible to get an idea on the reasons for the variations taking place in the structure of the barometric and thermal fields of the troposphere and of the lower stratosphere. From the structure of the isohypses and isotherms, as well as from the distribution of the wind velocities along the currents, the possibility of the appearance of new barometric formations, of their development, as well as the paths of existing cyclones and anticyclones are determined. The maps are used for calculating vertical air motion for determining favorable conditions for the development of cloudiness and the formation of atmospheric precipitation.

Barometric-topography maps for weather forecasting were prepared for the first time in the USSR in 1936 by the authors of this book. Their systematic preparation began on 25 November, 1937 at the Central Forecasting Institute. Since then they became extremely useful, not only for use in the Central Forecasting Institute, but also in all the Departments of the Weather Service.

Barometric-topography maps made it possible to pass from analysis of the surface weather map on a plane, to a spatial study of the lower atmosphere and of the processes developing there. Many assumptions made by scientists at various times were verified and improved in this way. They helped clear up many misconceptions about atmospheric processes, which were usually dogmatically expounded in textbooks on general and synoptic meteorology published until the 1950's.

We discussed absolute barometric-topography maps in Figures 33, 34, 69, 70, 74, and 75, and relative barometric-topography maps — in Figures 39–42.

Auxiliary maps and diagrams

In addition to the basic surface and high-level weather maps, auxiliary maps, e.g., peripheral maps, maps of isallobars, isallohypses, isallotherms, vertical currents, humidity, precipitation, distribution maps of ice-covered ground and hoarfrost, etc., may be prepared.

Peripheral maps are surface weather maps which are drawn from data of observations carried out between the basic times in a given place. They contain data on the weather from a denser network of meteorological stations than the basic weather maps. To give all these data, large-scale maps are usually used.

On isallobar maps the pressure variations at various points at ground level during some time interval are plotted, for example, during 12 or 24 hours. Isallohypse and isallotherm maps are drawn similarly.

On the maps of the vertical currents the calculated mean ascending and descending air motion during the period chosen (12 hours) are plotted. The humidity maps are prepared for various levels. They contain data on the moisture content of the air. On the precipitation maps data on the amount of precipitation that fell on the territory of interest during 12 or 24 hours and during other time intervals, as well as data on the maximum and minimum temperature, soil humidity, height of the snow cover, etc., are plotted. On the glazed frost and hoarfrost maps data on the rate of their deposition on cables and other objects during the observation time are plotted.

In addition to the basic and auxiliary maps, the meteorologist has at his disposal various diagrams and graphs. Aerological diagrams contain data on the variation of the temperature and humidity with height. Since on this diagram dry- and wet-adiabatic lines are represented graphically, by drawing the curve of the temperature variation with height, it is easy to determine the character of the atmospheric stratification, i.e., the degree of its stability or instability.

Figure 82 shows a somewhat simplified aerological diagram up to the 400-mb level. The diagram gives distribution curves of the temperature with height in the case of unstable (curve a) and stable (curve b) air equilibrium.

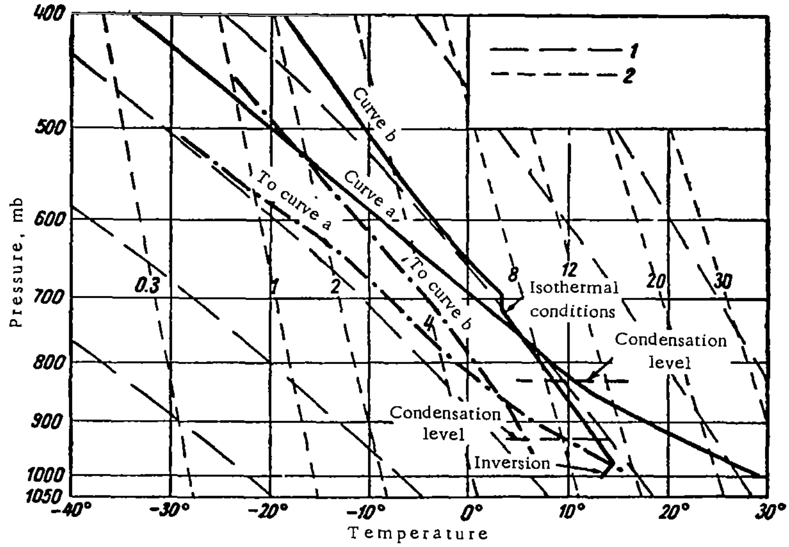

FIGURE 82. Aerological diagram with curves of the temperature variation with height

Curve a – the air is in an unstable state; curve b – the air is in a stable state; 1 – dry-adiabatic lines; 2 – curves of equal specific humidity.

In the first case (curve a) the temperature curve passes to the left of the dry- and moist-adiabatic lines. In other words, the lapse rate of the air up to saturation, i.e., up to the condensation level, is higher than the dry-adiabatic rate, and above this level, higher than the wet-adiabatic rate. Close to ground level temperature inversion, and higher up – isothermal conditions are observed.

In the second case (curve b) the temperature curve in the entire layer passes to the right of the dry- and moist-adiabatic lines. This means that the lapse rate up to the condensation level is lower than the dry-adiabatic lapse rate. Close to ground level temperature inversion, and at the 700-mb level – isothermal conditions are observed.

In addition to aerological diagrams, vertical cross sections of the atmosphere in the direction of interest are drawn. On the vertical cross section (Figure 83) isotherms (solid thin lines), fronts (thick inclined lines), cloudiness zones (shading), zones of aircraft icing and bumpiness, distribution of the specific humidity (dotted lines) and of the lapse rate (broken

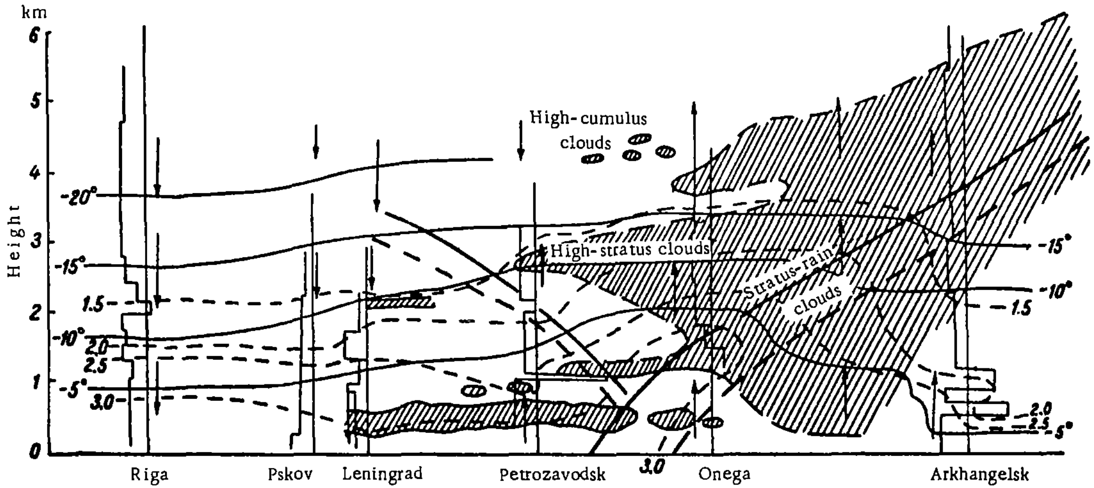

FIGURE 83. Vertical cross section of the atmosphere along the line Riga-Leningrad-Arkhangelsk, 15 December, 1952

lines along the vertical), ascending and descending air motion (indicated by arrows) are drawn according to data of aerological observations. Other weather data of interest to the meteorologist are also given.

The vertical cross section of the atmosphere on the section Riga-Leningrad-Arkhangelsk was drawn from sounding data of the atmosphere at points situated along the route, and from data of an experimental aircraft flight between Leningrad and Arkhangelsk. An occlusion front can be seen here, and behind the front, between Leningrad and Petrozavodsk, straticumulus clouds of small vertical thickness. Above 1 km there is almost no cloudiness. A different picture is observed in the section between Petrozavodsk and Arkhangelsk. There, in the region of the warm front, thick stratus-rain and high-stratus cloud forms (shading) were formed due to the ascending air motion. Over Arkhangelsk and other points temperature inversion was observed.

Short-term weather forecasting

Each weather forecast is preceded by multisided investigation of the atmospheric processes, carried out on the basis of data of ground and aerological observations. For this purpose use is made of surface and high-level weather maps, of aerological diagrams, of vertical cross sections of the atmosphere in the direction of interest and other auxiliary data. Simultaneous study of all the observation material at the disposal of the forecasting meteorologist makes it possible to get an idea about the spatial structure of the atmosphere and the weather over large geographic regions. By comparing in historical order weather maps, aerological diagrams, and vertical cross sections over a series of observation periods, the reasons for the variation of the atmospheric processes and weather on the preceding days are established. The meteorologist is interested in the development of processes, not only over the territory for which the forecast is prepared, but also far beyond its boundaries, since in one day air masses carrying good weather or cloudy weather with precipitation may traverse systems of anticyclones and cyclones – a distance of 1000 km and more.

If the processes remained unchanged, then it would be sufficient to determine the direction and speed of the air masses carrying some kind of weather. In reality, however, the atmospheric processes and the weather determined by them vary continuously. Atmospheric processes and the weather are influenced by the geographic environment, since the atmosphere and the underlying surface is in continuous heat and moisture exchange. This exchange, as we saw, is determined by the character of the underlying surface, by the temperature difference between the air and the underlying surface, by how far the air is from the state of saturation, by the evaporation conditions, etc. An important role is played by arising high-level frontal zones and their development.

The atmospheric circulation provides air exchange between low and high latitudes, as well as between continents and oceans as a result of which air masses of different properties approach each other (for example, cold and dry with warm and moist). Frontal zones are formed with their related cyclones and anticyclones, constituting the mechanism of interlatitudinal air exchange. Continual attention is paid by the meteorologist to frontal zones in which large energy reserves are concentrated and the air circulation in them is particularly intense.

We already know that vertical motion plays a very important role in the formation and variation of the weather, although the horizontal air transport is hundreds of times faster than the vertical transport. For example, over the European territory of the USSR during winter the mean horizontal transport speeds are equal to 30–40 km/hr, and the ascent and descent of air (in the absence of convection) takes place with a mean speed of 100–150 m/hr. Consequently an air mass moving with the mean velocity traverses about 800 km in 24 hours. During this time interval it ascends or descends a distance of 2.0–3.5 km. As a rule, due to the ascent or descent of the air by this amount, the weather in the moving air mass varies more than it would due to a horizontal displacement of even 1000–2000 km.

The sharpest weather variations at a point or region are observed upon the passage of atmospheric fronts. Since a particularly intensive horizontal and vertical air exchange takes place in the zone of fronts, the formation and scattering of the clouds takes place rapidly there. When a front approaches, clear and calm weather usually changes to bad during a few hours. Conversely, due to the activity of a front, overcast and rainy weather may begin to clear up as soon as the front moves beyond the boundaries of the given point or region.

Before preparing a weather forecast, the forecaster analyzes the atmospheric processes and the weather in the territory of interest. On the basis of thermodynamics and dynamics, he determines the character of and reasons for the variations taking place in the processes and the weather determined by them by means of weather maps. Making use of a whole series of assumptions and rules based on physical regularities, taken into account mainly qualitatively, the meteorologist determines the most probable variations in the existing barometric formations, and the direction and speed of their displacement. He determines the evolution and motion of atmospheric fronts and the cloudiness and precipitation zones connected with them, the circulation of cold and warm air masses and the variation in their thermodynamic properties due to the interaction of the air with the underlying surface, to vertical motion, etc. Then surface and high-level maps of the future pressure field are drawn with the expected positions of fronts, barometric formations, cloudiness, and precipitation zones.

Thus, the accuracy of a forecast depends first of all on the knowledge of the physical regularities of the processes developing in the atmosphere, on the amount and quality of the data of the ground and aerological observations, and to some extent on the working experience of the forecasting meteorologist under the given geographic conditions. The latter is due to the fact that the topography appreciably distorts the general picture of the processes taking place. For example, in hollows and valleys the air stagnates, and in clear weather as a result of the night radiation its temperature drops considerably more than over the neighboring elevations. During the transitional seasons this leads to frosts. Local conditions modify the direction and speed of the wind, forms fogs in low places, intensifies or weakens precipitation on the windward or lee slopes of mountains respectively, etc.

The general weather forecast consists of forecasts of each element or weather phenomenon; this produces certain difficulties. For example, in temperature forecasting one uses the characteristics of the air mass which should be over the given region during the period of the forecast. Thus the assumed variations in the initial temperature along the path of the air depending on the properties of the underlying surface, on the cloudiness, turbulence, etc., must be determined.

The forecasting of continuous rains is based on the forecasting of the motion of atmospheric fronts and the cloudiness and precipitation zones connected with them, as well as of their expected evolution. Forecasting of snowstorms requires a knowledge of the future air temperature, wind velocity, and precipitation. To forecast glazed frost and its intensity one has to know not only the future air temperature at ground level, but also the future distribution with height, the wind speed and direction high up, as well as the moisture content of the air and the precipitation.

Weather forecasts are comparatively easy to prepare and have a high accuracy in those cases when sharp variations do not take place in the development of the synoptic processes. In those cases when the fields of the temperature, pressure, and prevailing currents in the atmosphere rearrange with the associating appearance and development of atmospheric fronts and barometric formations, the meteorologist encounters great difficulties in preparing the forecast. Unsuccessful weather forecasts are mostly due to difficulties in determining the rapidity of the variations of the synoptic processes.

In weather forecasting it is most important to determine the variation of the atmospheric circulation, which is closely related to the problem of forecasting the pressure field. Over the last ten years, both in the USSR and abroad, the theory of pressure variation – the basis of the methods of weather prediction – has been intensively developed.

Already at the turn of this century, individual scientists (Bjerknes, Richardson, Angerovo) tried to predict the weather. However, at that time no practical use was seen in the prediction, since weather forecasting took many days of complicated mathematical calculations. There were also skeptics who said that weather prediction was impossible.

Soviet scientists proved the practical possibility of predicting the elements characterizing the weather. Important successes in this field were achieved in 1940. Applying the equations of motion of a baroclinic fluid to the atmosphere, I.A. Kibel' showed how in principle the operational prediction of the pressure, temperature, and wind a day ahead was possible. Of course the solution of such a complicated problem was far from perfect, since important simplifications were made in the solution of the equations. The accuracy of these forecasts was lower than the accuracy of synoptic forecasts prepared by qualitative methods, which until today are the basic methods all over the world.

In the USSR, the USA, and other countries in the subsequent years, the calculational methods of forecasting the pressure, temperature fields, etc., were improved. Instead of forecasting for the middle level (3–4 km) the problem was solved for a number of levels. In these cases there is so much calculation that it is practically impossible to manage without computers.

Fast computers were used for the first time in 1950 in the USA. They are now used for weather forecasting in the USSR, the USA, Switzerland, England, and other countries. The equations used for calculating the variation in the atmospheric pressure make it possible to obtain the pressure field every 12 or 24 hours from the initial pressure distribution over a considerable territory, i.e., from the current weather map.

However, as was said above, there are still many difficulties in working out perfect methods of weather prediction. These are due to the fact that a large number of complexly interacting factors are involved and therefore one has to resort to simplifications in the mathematical solution of the problem.

Nevertheless, the modern methods of weather prediction gradually improve. There is no doubt that the perfection of weather forecasting methods depends on developing the theory of pressure variation and using fast computers.

Long-range weather forecasting

Long-range weather forecasts of a month or a season are of particular importance to the national economy. They are important for planning the work of industry, transport, and particularly agriculture. They are necessary for the preparation of agrometeorological and hydrological forecasts (the latter include forecasts of spring high water, rate of spring flow, and the freezing and thaw of rivers and of large and small water reservoirs). The methods of long-range weather forecasting are less developed than the others. Among the most important and difficult problems we have to deal with today (e.g., the problem of atomic energy), we can include the problem of long-range weather forecasting.

The Soviet Union is the home of long-range weather forecasts. The first practical steps in this field were undertaken at the Central Physical Observatory in 1915 by the distinguished meteorologist B.P. Mul'tanovskii. Systematic preparation of long-range weather forecasts began in 1922. Mul'tanovskii's method was developed by S.T. Pagava, N.A. Bulinskaya and others.

Simultaneously with the further improvement of this method during the last 10–15 years, other methods were also developed. Scientists paid special attention to high-level weather maps – barometric-topography maps. These maps were used by G.Ya. Vangengeim and A.A. Girs for standardization of large-scale atmospheric processes in order to establish the regularities of these processes in the whole northern hemisphere so that long-range weather forecasts can be prepared.

In the USSR long-range weather forecasting is prepared for periods of from several days to a month and season. The weather forecasting for several days is based to a certain extent on an artificial breaking down of the synoptic processes into periods of 5–7 days. During this time interval over a relatively large territory (for example Europe and western Siberia), the character of the atmospheric circulation is maintained approximately constant. Although the weather in individual sections of this territory may vary considerably during the period, its forecasting is not so difficult, since the basic synoptic processes vary insignificantly during the period. When there are changes in the synoptic processes, weather forecasts are much more difficult to prepare.

The preparation of a weather forecast for a month can be divided into the following operations:

1) Atmospheric processes similar to those of the previous season are chosen.

2) Some "reference" processes are determined in the month being forecast (by means of the rates and phases of development of the processes over a large territory, for example, the northern hemisphere or Eurasia).

3) Analogs of the months being forecast are chosen from the number of analogs chosen for the previous season.

4) A sequence scheme of the development of the processes in the months being forecast is set up.

5) The text of the forecast is prepared.

In contrast to short-range weather forecasts, forecasts for a month or a season give a more general characterization of the weather. Indications of the deviation of the assumed monthly amount of precipitation from the norm are also given. In addition, on the basis of reference positions in the monthly and seasonal forecasts indications of the expected sharpest weather variations are given as well as a general characterization of the weather.

Due to the considerable difficulty in allowing for even the most important factors determining the weather variations, long-range forecasts naturally have a lower accuracy than short-range ones. In particular, modern methods of long-range weather forecasting cannot take into account the influence of the heat inflow on the circulation of the atmosphere, although it is known that this cannot be neglected when determining the character of the circulation during a month, and all the more so during a season. Unfortunately, since quantitative forecasting of the heat inflow is very complicated, one has to be satisfied with various empirically established relationships and use them in the preparation of long-range weather forecasts.

Sharp weather variations

The weather in middle and high latitudes very often undergoes sharp variations. An example is the sharp weather variation which took place on 27–28 October, 1952 in the south of the European territory of the USSR.

During 27 October, in the eastern Ukraine and in adjacent regions, clear weather prevailed with very weak winds, night temperatures of about 6°–8°C, and day temperatures of 12°–15°C above zero. Due to peculiarities in the atmospheric circulation, in the evening of 26 October a cyclonic disturbance arose over the region of the Hungarian lowlands. All conditions existed there for its violent development, i.e., for the deepening and for the extension of the activity territory. The most important of these conditions were advection of cold air to the tail of the arising cyclone, intensification of the ascending motion of the air, and a pressure drop in its system. Cloudiness developed, and much precipitation began to fall; in one night it amounted to 10 mm.

Air transport from west to east over the cyclone prevailed. According to the established rule, cyclones and anticyclones move approximately in the direction and with the speed of the air currents prevailing in the middle troposphere above them. The cyclone therefore had to move toward the east with a speed of about 50 km/hr. Consequently, traversing about 1200 km in 24 hours, it should have appeared a day later over the east of the Ukraine.

Indeed after 24 hours the pressure field and the character of the weather appreciably changed on the south of the European territory of the USSR.

The cyclone also changed appreciably: this was no longer a recently generated small disturbance, but a barometric formation of large diameter with strong and gusty winds. The pressure at its center reached 995 mb, i.e., the cyclone deepened by 15 mb as compared with the past days. Such large pressure variations take place in violently developing barometric formations. Simultaneously with the deepening, the weather continued to worsen in the whole system of the cyclone, the vertical thickness of the clouds increased, etc.

The bad weather moved in the direction of the prevailing air currents through the whole Ukraine. As a result the clear, calm, and warm weather changed to bad. As long as the cyclone was passing, continuous precipitation fell. However, a day later, after it moved to western Siberia, cold air masses penetrated in the tail of the cyclone from the north giving rise to a considerable drop of temperature in the south of the European territory of the USSR associated with clear weather and a small amount of clouds.

We described here only one sharp weather variation, caused by the development and movement of a cyclone. Similar cases can be observed daily in various regions of the northern and southern hemispheres.

WEATHER FORECASTING ACCORDING TO LOCAL CHARACTERISTICS

Man's dependence on the weather forced him to observe its variations and find criteria by which it could be predicted. Some of the popular criteria for predicting the weather have been justified scientifically. Most of them, however, have no scientific explanation, since they have nothing in common with the development of atmospheric processes.

A list of criteria and popular proverbs on weather changes can be found in many sources in the literature. Here we acquaint the reader only with those criteria or indications of future weather variations which have physical grounds. The reader can always verify them in practice and use them in cases of need.

In the age of radio and television the interest in local criteria of weather variation gradually decreases. The reason for this is that the scientific weather prediction has a high accuracy. Thus, for example, the accuracy of short-range weather forecasts exceeds on the average 80%. In other words, out of ten weather forecasts prepared, only two fail. Such an accuracy is no doubt higher than the accuracy of forecasting by any set of local criteria. Moreover, weather forecasts prepared by the weather service are widespread, since radios can be found in any corner of our country, and consequently, one can listen to the announcement of the future weather in one's own region. However, the weather forecasts transmitted by radio are usually prepared for a large region. The knowledge of local indications, based on noninstrumental observations at one point, may in many cases help to improve the weather forecasting for this point.

Let us see which of the weather elements can be used to get an idea of future weather variations. At the same time we shall determine the basic types of weather which are convenient for forecasting by local criteria.

Principal criteria of weather variation

Some idea of future weather variations can be obtained from the clouds, wind, optical phenomena, etc. The form and height of the clouds, the direction and speed of their motion, as well as the variation of these cloudiness characteristics with time, are of great importance. From observations of the cloudiness one can determine the state of the atmosphere higher up, not only over the observation point, but also at some distance from it.

From optical phenomena such as light circles around the Sun and Moon (halo), and rings about the Sun and Moon (corona), one can determine the structure of high-lying clouds. From the character of the dawn one can get an idea about the humidity and the presence of various impurities in the air.

The approaching change in the weather directly over the observation point is determined from the variations in the temperature, and in the direction and speed of the wind. The weather variation is also indicated by the character and duration of precipitation, by thunderstorm phenomena, horizontal visibility, etc.

One criterion is not sufficient to determine an approaching weather variation. One should try and account for all of them. To be able to use the criteria correctly, the weather variations under the given physicogeographic conditions must be followed and determined accurately every day. It is natural that the greater the experience in observations of weather changes and the more correctly the criteria used are explained, the more successful will be the weather predictions.

Basic types of weather

It was indicated above that the weather at any point of the terrestrial globe, particularly at middle and high latitudes, undergoes continuous variations, depending on a large number of factors. These variations are so diverse, that, strictly speaking, the weather features observed during one day are rarely repeated in all its details on any other day. But a similar type of weather is often observed. For example, dry weather with a small amount of clouds is often observed during the year, although the individual features of the cloud development, of the variation of the temperature, humidity, and pressure of the air, and of the wind and cloudiness at a given point may be quite different. Of course, the concept of similarity is relative, and consequently depends on the conditions assumed.

When forecasting the weather from local noninstrumental observations, the diverse types of weather can be divided into the following four types:

The first type – clear weather or weather with a small amount of cloud and no precipitation.

The second type – cloudy weather which clears up with short torrential precipitation.

The third type – overcast weather with low cloudiness.

The fourth type – bad weather with continuous precipitation.

Each of these weather types depends on the general character of the development of the atmospheric processes and on the activity of fronts, cyclones, and anticyclones over the territory under consideration.

1. Clear weather or weather with a small amount of cloud and without precipitation is usually characteristic of anticyclones. During the winter it sets in after a sharp cooling and gradual weakening of the wind. Frosty weather with weak winds and calm is often observed in Siberia during the winter. In this case clouds are not formed as a rule, since at an air temperature of -20° to -30°C the moisture content of the air is negligibly small. For precisely this reason such weather is observed not only with anticyclones, but often even with cyclones. Over the European territory of the USSR, on the other hand, clear weather or weather with a small amount of cloud is rare during the winter months, since not only with cyclones, but also with the western half of anticyclones low-level cloudiness and fogs form due to the transport of warm and humid air from higher up in the south.

During the summer clear weather or weather with a small amount of cloud is most often observed in places where, due to the strong heating of the air and the low evaporation, the air is removed from the saturation state. This no doubt contributes to the formation of anticyclones, in the system of which the air is adiabatically heated as a result of its descent. But during the summer sometimes even in cyclones there is clear weather or weather with a small amount of cloud, if the moisture content of the air is low and its relative humidity does not exceed 20%–40%. Thus, for example, over the Lower Volga and the Urals cyclones sometimes do not give precipitation during summer, although the air ascends and does not descend there. Moreover, over North Africa (Sahara), Arabia, Iran, and Central Asia a low-pressure region exists almost throughout the entire summer and precipitation does not fall; this again is due to the low relative humidity of the air.

2. Unstable, cloudy weather which clears up with short torrential precipitation, often associated with thunderstorms, is typical of the summer season. Unstable weather, however, is sometimes also observed during winter in the tail of cyclones, behind a cold front. Over the continents such weather is formed in those cases when the cold air arriving from other regions becomes unstable due to the heating of its lower layers over the warm underlying surface. The appearance of air instability leads to the development of convective vertical currents, as a result of which cumulus and cumulonimbus clouds are formed. The latter attain a very large vertical thickness and, depending on the air temperature, give torrential rain or snow. Torrential precipitation falls more often during the winter over the oceans. Its formation is due to the arrival of cold air from the cooled land to the warm ocean surface and the warming and moistening of its lower layer.

This type of weather is usually characterized in observations at one point by a sharp transition from clear to cloudy weather followed by torrential precipitation; then breaks appear in the clouds and it clears up again. The increase in the cloudiness, the fall of torrential precipitation, and the rapid clearing up are related to the passage of a cold front through the given point. However, when the cold air behind the front is unstably stratified and its moisture content sufficient, thick cumulus and cumulonimbus clouds also appear inside this relatively uniform cold air. Since these convective clouds usually do not form a single cloud mass, the cloudiness increases or decreases, and at times even clears up over the observation point. But since the air is heated from the underlying surface, unstable weather over land is typical of middle latitudes during the warm half year. To the south of our country, in the Republics of Central Asia and Transcaucasia, as well as over the north Caucasus, Lower Volga, Ukraine, and Moldavia, torrential precipitation is the main part of the precipitation not only during summer, but also spring. During winter on the continents of the northern hemisphere unstable weather with torrential precipitation is more often observed at latitudes of 30°–40°, since there the air is intensively heated and instability develops. During spring and summer unstable weather is observed in the northern regions, where at times it replaces cloudy weather by continuous precipitation.

As was said above, the prediction of torrential precipitation for an individual point is very difficult, since it does not fall everywhere as frontal precipitation, but only here and there. A single cumulonimbus cloud gives rain only in that region over which it passes, and therefore, within the boundaries of a large town, torrential rains fall only in certain places.

3. Cloudy weather with low-level strati cloudiness, from which light precipitation often falls, is usually observed during the winter. The formation of this type of weather is connected with the motion of warm and humid air over a cold underlying surface. During winter such weather is often observed over the western half of the European territory of the USSR, where the western periphery of a powerful anticyclone is situated. In these cases, relatively warm and humid air moving from south or southwest comes into contact with the cold underlying surface and cools down, as a result of which it becomes saturated and fog or low-level stratified cloudiness is formed. The height and vertical thickness of these clouds are usually small, since their formation takes place during the weak vertical ascent of the air which encompasses the lower layers of the troposphere.

Cloudy weather with weak precipitation exists also in a zone of eroded or unclearly displayed atmospheric fronts and in the warm sector of young cyclonic disturbances, where weak ascending air motion is observed.

4. Bad weather is typical of zones of cold and warm fronts of developing cyclones, when clouds form very quickly and precipitation falls. This weather type is characterized by strong and moderate gusty winds. The fall of snow is associated with heavy snowstorms. Such weather begins upon the approach of a cyclone to the observation point and is maintained as long as the cyclone and its related system of atmospheric fronts is situated over the given region. The frequency and the duration of bad weather are connected with the active cyclonic activity, and therefore bad weather is more often observed during the winter in the northern and central regions of the European territory of the USSR and of western Siberia and more rarely in the southern regions of the country.

In the previous chapters we dealt with the complexity of atmospheric processes and the difficulties of predicting the weather. Using the appropriate equipment, the meteorologist at the weather bureau studies the direction and speed of motion of barometric formations and of the atmospheric fronts connected with them, determines the character of the variations which took place in the processes and in the weather during the previous days, and prepares the weather forecast for the region or point of interest.

An observer, not having weather maps at his disposal, is not able to determine accurately the paths of cyclones and anticyclones, of atmospheric fronts, or of cold and warm air masses, nor their qualitative variations. In local weather prediction use is therefore made mainly of those criteria from which some idea about the movement of the barometric formations and fronts can be obtained.

Criteria for the arrival of clear weather or dry weather with a small amount of cloud

Bad weather is often replaced by clear and dry weather. The main criteria for this change are the gradual weakening of the wind, and the decrease in the cloudiness until it clears up. The improvement in the weather is associated with a rise in the atmospheric pressure. During the winter the temperature drops appreciably, during the summer, it rises.

Usually the slower this change of weather, the more probable it is that the arriving clear and dry weather will be maintained for a prolonged time (several days). This is due to the fact that little-developed cyclones, alternating with barometric crests, rapidly interchange with each other on their path in any direction. The passage of a cyclone through the observation point is associated in this case with bad weather, and when the cyclone is replaced by a crest, clear weather or weather with a small amount of cloud begins.

If the weather change is slow, taking a day or more, it follows that the cyclone is a deep and powerful barometric formation. In the tail of a deep cyclone the penetration of cold air masses is intensive. The strong rise of the atmospheric pressure in the tail of a cyclone, which has moved beyond the boundaries of the observation point, is associated with the appearance and intensification of a powerful anticyclone; this nearly always remains almost immovable for 2–4 days. Clear and dry weather is maintained until the formed anticyclone collapses and goes away.

Cyclones and anticyclones mostly move from west to east. A criterion for a possible improvement in the weather is therefore a variation of a southeasterly and southerly wind into a northerly or northwesterly one and a gradual weakening of its strength.

During clear weather or weather with a small amount of cloud, the daily march of the temperature, humidity, and wind is clearly displayed in the warm half of the year. During winter the daily march is displayed appreciably weaker. At night, due to the radiation of the surface of the Earth, the temperature drops.

In the transitional periods of the year, during spring and fall, the air temperature often drops to below 0°C and frosts begin due to the radiational cooling; when the temperature drops, the relative humidity rises. If the moisture content of the air is sufficient, saturation begins and the water vapor condenses on the plant cover, roofs, and other objects – during the summer in the form of dew, during the winter in the form of hoarfrosts.

Often, particularly over lowlands, fog which does not have a large vertical thickness is formed and scattered by the first rays of the Sun. The earlier in the evening that the fog appears, the later in the morning it is dispersed.

By day the stratification of the lower air layers becomes unstable due to the heating. By night, on the other hand, due to the cooling of the air from the surface of the Earth, surface temperature inversion appears. In lowlands, therefore, it is colder than in higher places.

Due to the nonuniform heating of the air by day the wind velocity somewhat increases and slight gustiness is observed. Toward evening the wind weakens appreciably.

During clear weather in mountainous countries the mountain-valley air circulation is well displayed, and on coasts of seas and lakes – breezes are observed. We recall that in mountains the wind is directed from the valleys to the mountains by day, and by night – from the mountains to the valleys. On shores of seas and lakes the wind is directed from the water surface to the coast, by day and at night – in the opposite direction. The more unstable the air, the stronger the mountain-valley and breeze circulation.

One of the main criteria for stable and good weather is the daily march of the cloudiness. After a clear night, the lower air layers are heated and the night temperature inversion is destroyed as the Sun rises. As the air stratification becomes unstable, ascending currents appear and at 0900--0100 hours cumulus clouds are formed. If at midday the clouds do not develop upward, have flat bases, and are sparse, then toward evening they spread out and clear weather arrives again.

The amount of clouds and their vertical thickness may also be considerable, but if at midday their growth stops, then in the evening they scatter. However, in such cases, cumulus clouds scatter more slowly and their tops may be transformed into melting straticumulus clouds in the second half of the day.

Insufficient development of cumulus clouds is due to two reasons. The first is that owing to the low moisture content (high dryness) of the air the condensation level with which the lower rim of the clouds is connected is situated so high up that the instability energy existing in the atmosphere is insufficient to lift large air masses considerably higher than this level and cause violent cloud formations. The second reason is the insufficient thickness of the unstably stratified air layer. In this case the air above 3-4 km is usually in a stable state. Ascending convective currents developing in its lower layers, encountering therefore an obstacle in the form of a stably stratified air layer, decay and the cloud formation does not extend upward.

As we shall see below, when the moisture content and the instability are sufficient, a thick cumulus and cumulonimbus cloudiness develops and torrential precipitation associated with thunderstorms falls.

One of the criteria for continuous weather with weak winds is radio disturbances which effect radio receivers. When the atmosphere is unstable electric discharges arise whose intensity depends on the degree of instability. Therefore, in the case of a weather with weak winds, the noise and crackle in radio receivers are so weak that they do not impair reception. Since at night surface inversion appears in the case of weather with a small amount of cloud, vertical air currents and their accompanying radio noises become weaker. By day, with the appearance of instability, the crackle in radio receivers is intensified. This also explains the well-known fact that the audibility during winter is better than during summer. When using this criterion, it should be remembered that bad reception of radio transmissions is also connected with processes in the ionosphere.

The color of the sky is also an indication of the preservation of clear weather or weather with a small amount of cloud. If good weather sets in after the penetration of little-dusted air masses from the northern quarter of the horizon, the sky in the morning is blue, by day whitish. The sunset has a golden or orange color, changing later to greenish-yellow.

As a criterion for the maintenance of little-cloudiness or clear weather one may use the variation of the atmospheric pressure read from an aneroid barometer. It has already been said that cyclones and anticyclones move continuously. The approach of a cyclone is associated with a drop and of an anticyclone with a rise in the pressure. In stable good weather the pressure does not undergo noticeable variations -- it has a small daily variation, dropping by day and rising by night. This means that the pressure field in the region of the observation point remains almost unchanged. An upsetting of the daily pressure march foretells a change in weather.

Criteria for cloudy weather with clearings up and short torrential precipitation

This type of unstable weather is caused by the passage of a cold front, after which, as we saw, clear or little-cloudiness weather may begin. But in those cases when the cold air mass behind the front becomes unstable due to the heating of the lower layer, cumulonimbus cloudiness develops bringing torrential precipitation.

A rapid development of cumulus clouds in the morning is characteristic of unstable weather. The clouds gradually acquire a lead color, and in the process of their development they are transformed into cumulonimbus clouds whose tops spread out to large heights, often taking the shape of an anvil. In this case a shower associated with a thunderstorm should be expected. If toward evening it does not become cooler, the showers with thunderstorms may also occur at night. When it does get cooler at night the cloud development stops.

At middle latitudes the height of thunderstorm clouds reaches 5–7 km. The higher the thunderstorm clouds develop, the more intense the shower. When their vertical thickness is considerable, the fall of hail should be expected. During winter snowflakes fall from forming cumulonimbus clouds, during spring — soft hail with large snowflakes or raindrops.

Over mountains cumulus and cumulonimbus clouds cover the windward part of the mountain regions and individual tops particularly intensively.

If the cumulus clouds cover a considerable part of the sky, but their tops do not extend upward in the form of towers but rather spread, then a shower may not be expected.

When conditions exist for the formation of shower clouds and precipitation, the sky in the clear gaps between the clouds has a blue color. When conditions for the fall of showers do not exist, the sky in the clear gaps between the clouds is whitish.

For shower clouds to form, in addition to the heating of the lower air layers and development of unstable stratification, there must also be sufficient moisture content. The higher the moisture content of the air, the higher the probability of showers and thunderstorms. Therefore, prior to a thunderstorm a stuffiness is usually felt, being a result of the high humidity of the air.

Night fogs arise after daytime thunderstorms in cases of high air humidity. Fogs are not formed before a night thunderstorm.

Cumulonimbus clouds are not stable formations. They usually spread, transforming into straticumulus, high-cumulus and cirrus clouds. In strati clouds one often sees light rings (corona) around the Sun, and in cirrus clouds — yellowish rings (halo).

A red sunset indicates the possibility of a night thunderstorm.

Unstable weather may last for several days. But usually the intensity of the showers gradually decreases.

Criteria for the maintenance of overcast weather with low-level cloudiness

Overcast weather is characterized by complete coverage of the sky, by low-level clouds of stratified forms, by a small daily variation of the temperature, and by an almost unchanging direction and speed of the wind. Since the formation of continuous stratified cloudiness takes place upon the motion of warm air over a relatively cold surface of the Earth, the air temperature rises when there is a change to overcast weather.

In such a case the wind usually gradually intensifies, becoming moderate. Winds prevail in the western and southern parts of the horizon.

When strati clouds develop light precipitation falls. Strati clouds often descend to ground level, forming a fog which may last for several hours and even days.

The air pressure varies only little. When the pressure sharply rises or drops, it indicates a change in the type of the weather.

Criteria for the arrival of bad weather

The transition to bad weather is associated with an upsetting of the daily march of the meteorological element. The latter usually varies sharply. This is connected with the rapid displacement of atmospheric fronts, cyclones, and anticyclones.

We recall that fronts are most sharply displayed in the system of developing cyclones. A worsening of the weather is therefore connected with the approach of a cyclone to the observation point; this may be inferred from the appearance of the corresponding clouds (see Figure 54).

FIGURE 84. Cirrus clouds

One of the indications of weather worsening is the coloration of the sunset. This criterion can be used in many cases, since warm fronts usually move from west to east. If the zone of continuous precipitation moves from the west, then the sunset has a red, sometimes purple-red color in clear weather. In this case the certainty of the approach of bad weather is stronger if on the day before the sunset was yellowish, orange, or rose in color. From the sunset it is possible to get an idea about the weather at a distance of up to 1000 km from the observation place.

Cirrus clouds are correctly considered as the forerunners of bad weather. On a clear day strips of transparent cirrus clouds, well discernable at sunset, appear on the horizon (Figure 84). As they approach they become denser, forming semitransparent calm strips which gradually extend over the whole sky and pass into a cirrus-strati form. The faster the clouds move, the sooner does the weather worsen. Their appearance above the observer indicates that bad weather is at a distance of 200–400 km from the observation point at that side of the horizon from which they come. Consequently, with cloud velocities of about 30–40 km/hr cloudy weather with precipitation will reach the observation area in 7–10 hours.

As the cirrus clouds become denser light circles with a yellowish shade (halo) appear around the Sun or Moon. The faster the light circles disappear, the faster the bad weather arrives. The disappearance of the circles around the Sun indicates the appearance of middle-level clouds, i.e., of high-stratus clouds through which the Sun appears as a faint circle. With the appearance of denser black-grey stratinimbus clouds precipitation begins to fall.

When observing the variation of cloud forms and the motion of clouds attention should be given to the direction of motion of clouds at different levels. Thus, for example, if cirrus clouds deviate to the right of the direction of motion of lower lying clouds, this may serve as an additional indication of the approach of a warm front and its characteristic weather. The greater this deviation, the faster the weather worsens.

One should not confuse frontal cirrus clouds with cirrus clouds appearing as a result of the spreading of the tops of high-cumulus clouds. The difference is that frontal cirrus clouds move from one direction of the horizon and in addition they are appreciably denser. Cirrus clouds appearing as a result of the spreading of high-cumulus clouds, usually appear disorderly in various parts of the sky.

It should be realized that the appearance of cirrus clouds is not always associated with a subsequent worsening of the weather, even if they are frontal ones. This is due to the fact that the direction of motion of a cyclone and its related front often varies in accordance with the development of atmospheric processes. Therefore, in spite of the appearance of cirrus clouds at the observation point the main frontal cloudiness and continuous precipitation may appear wide off the given point due to the variation in the direction of motion of the front.

Cases when an intensive atmospheric front is initially eroded and, although the cloud system remains, the clouds spread and do not give precipitation are also frequent. Cirrus clouds, therefore, do not always foretell the arrival of bad weather. The erosion of warm fronts occurs more often during the summer over the European territory of the USSR and to the east of it.

In order to make less mistakes when forecasting weather from local observations, other criteria indicating weather worsening must also be taken into account.

At middle latitudes bad weather is typical of the winter half year when the fronts and cyclones are intensive and their activity prolonged.

During winter, the worsening of the weather is associated with a rise in temperature. During summer, on the other hand, hot weather becomes warm and often even relatively cool.

The wind intensifies as a rule. Northeasterly and northerly winds usually change into westerly, southwesterly, and southerly ones. When there is a snow cover and snow falls, snowstorms are observed.

The air humidity gradually rises and the pressure drops, indicating the approach of a cyclone.

If bad weather is connected with a sharp cold front, then the air pressure before the worsening of the weather slowly drops, and then usually rises sharply.

When a thunderstorm front approaches, clouds of all levels appear almost simultaneously. First lenticular clouds appear, followed by all the remaining cloud forms. The wind then intensifies sharply, and the temperature drop is associated with a pressure rise— after several hours the weather clears up (see Figure 58 b).

Bad weather connected with the passage of a cold front is usually observed for quite a short time interval; afterwards the weather improves.

THE POSSIBILITIES OF MODIFYING THE CLIMATE AND THE WEATHER

If a timely and correct warning of expected dangerous weather phenomena can be given, then the necessary measures to reduce the damage that they can cause can be taken. It is natural that man will not be satisfied with weather forecasting alone, which is not always accurate, but will try to influence the atmospheric processes causing weather variations and thereby avoid the undesirable weather phenomena. In other words, cannot man, who has mastered so many of nature's secrets, master also the weather?

To be able to control the weather means to be able to control the air temperature, i.e., to moderate the heat in the desert where it is destructive to animals and plants, to bring more heat to the cold regions of the Earth, for example, the polar regions and tundra, etc. To control the weather means to create conditions for the formation of clouds and precipitation where they are necessary. To control the weather means to moderate the force of hurricanes and storms, to prevent the fall of hail, to eliminate thunderstorms and strong showers, snowstorms, dust storms, and, finally, such dangerous natural phenomena as droughts, dry winds, etc.

Is it in the power of man to control the weather?

We have already said that the formation of the weather and the climate is due to the three basic factors – solar radiation, underlying surface, and atmospheric circulation. The actions of these factors are closely related to one another.

The components of the radiant energy balance depend mainly on the latitude of the place, on the time of the year, and to a certain extent on the character of the underlying surface, which creates nonuniform conditions for the absorption and reflection of the solar radiation, and thus nonuniform albedo. The components of the radiant energy balance are also influenced by the circulation of the atmosphere, since the prevailance of overcast or clear weather over some part of the Earth, manifested both in the radiant heat exchange and in the moisture exchange, depends on the character of the atmospheric circulation. The underlying surface in turn is subjected to the influence of the solar radiation and of the atmospheric circulation, the latter depending on the distribution of the heat inflow over the terrestrial globe and on the character of the underlying surface.

The role of each of these factors in the formation of the climate and weather is so large that the ability to control or regulate the action of just one of them would make it possible to solve the problem to a certain extent. Thus if man could influence the atmospheric circulation and regulate it, the problem of weather control would be basically solved.

Let us attempt to clarify man's chances of influencing these factors.

1397

The Sun is far away from the Earth and man cannot influence the variation of the solar radiation. But, by modifying the character of the underlying surface, man can influence the variation of the components of the thermal balance.

The atmospheric circulation depends not only on the solar radiation but also on the conditions of the underlying surface, and therefore the circulation could be modified to some extent. This would, however, require too much energy.

The energy resources on Earth are enormous, yet man has always been searching for new kinds of energy. Until today, however, only energy reserves created by the Sun, since coal and oil are concentrated solar energy accumulated by plants during many millions of years, have been used.

Other energy forms are the hydroenergy of rivers and the energy of the wind. Hydroenergy is also created by the Sun, since an important role in the process of water rotation is played by the evaporation of water from the oceans and continents. The wind (air currents) is also a product of solar energy. Its appearance, as we saw, is due to the nonuniform distribution of the heat over the terrestrial globe.

All these kinds of energy, in spite of their huge reserves, cannot become a powerful source of influencing and modifying the character of the atmospheric circulation.

The world energy reserves grew considerably as a result of the mastering of the use of atomic energy. According to approximate calculations, the energy contained in uranium alone exceeds by several orders of magnitude the energy of the world reserves of coal and oil. The energy released in the fission of 1 kg of uranium is equal to the energy obtained by the burning of 20 million kg of coal.

However, atomic energy cannot yet be used to influence the circulation of the atmosphere on a large scale. Interesting data are given in a paper by E.K. Fedorov, published in 1954. It was found that during a single storm of ordinary force about as much energy is spent as is contained in 10–15 hydrogen bombs. In order to artificially cause a storm at sea, it would thus be necessary to explode approximately one hydrogen bomb per second.

To get an idea about the huge losses of energy in elementary atmospheric processes, we may look at two other examples, taken from a paper by E.K. Fedorov, published in 1960. With the appearance and development of several ordinary cumulus clouds in the course of 3–4 hours, about 30 million kwhr are spent. To produce this amount of energy in this amount of time several hydroelectric stations equivalent to the Kuibyshev station would have to be used. Another example: the energy of a wind blowing with a velocity of 20 m/sec along a front 200 km long is equivalent to several millions of kilowatt hours.

In other words, to influence the circulation of the atmosphere with the purpose of controlling the weather on a large scale, we need a powerful energy source. It is obvious that such a source could be the Sun, whose energy, as we saw above, is incomparably larger than the energy of other sources. But a considerable part of the solar energy arriving on Earth is uselessly radiated back into space, and the human genius has not yet been able to find methods for using it extensively.

In the chapter "The inflow of solar energy" we mentioned that man already finds ways for using solar energy, e.g., for obtaining substances rich in chemicals in order to transform the chemical into electric energy. But let us return to the possibilities of atomic energy.

It has now become possible to use atomic energy for industrial purposes. As a result of the efforts of Soviet scientists, the first atomic power station in the world, producing 5000 kw, already operates in the Soviet Union.

In our country and abroad, powerful atomic power stations are being constructed. It can be assumed that in the near future the greatest discovery of the 20th century – atomic energy – will be widely used throughout the world.

The world reserves of uranium and thorium amount to about 20 million tons. The energy of these two heavy elements is equivalent to the energy of tens of trillions of tons of coal. However, the extraction of these rare elements and the production of electric energy still have unpleasant aftereffects. Scientists are therefore concentrating on controlled thermonuclear reactions producing the atomic nucleus of the element helium from nuclei of light elements; this is accompanied by the release of a colossal amount of energy. For example, 1 kg of heavy hydrogen – deuterium – transformed into helium, releases an amount of energy equivalent to that obtained by burning 40,000 tons of coal. The explosive energy of a hydrogen bomb is the energy of thermonuclear fusion of helium from ordinary hydrogen. These processes, taking place on the Sun and in stars, are the source of their energy.

If it will become possible to realize controlled thermonuclear reactions, then the problem of new energy sources will be finally solved. In this case heavy hydrogen will become the new fuel. It suffices to say that 1 liter of water contains heavy hydrogen equivalent in energy content to 400 kg of petroleum.

It is hoped that atomic energy will be used to influence local processes of weather formation, for example, to intensify or stop precipitation, to prevent damage caused by hail, storms, showers, etc.

Certain successes have already been acheived in a number of problems connected with influencing the weather.

Meteorological processes are often in such a phase of development that a small interference from the outside is sufficient to modify their character. For example, if the air is saturated and clouds are formed, but precipitation has not yet fallen from them, then by introducing some substance (reagent) into the cloud the precipitation can be accelerated. In this case solid carbon dioxide (CO_2), finely-ground ice, and other substances are used as reagents. In particular, experiments are being made to prevent damage by hail: obviously it is possible to force a thunderstorm cloud to dispose of a part of its moisture before the water droplets and ice crystals transform into hail. Successful experiments are being made to disperse fog and low-level cloudiness at air temperatures below 0°C. The scale of these experiments is always being enlarged. However, many difficulties must still be overcome before it will be possible to affect atmospheric processes of even local importance.

Commercial organizations were set up in the USA for the purpose of producing "artificial precipitation." Wide-scale advertisement helped to attract innocent farmers into the venture, but they soon realized that it was a fraud.

There are also adventurers who played with the idea of a so-called "meteorological war." According to their plan the atmospheric processes have to be influenced in such a way that drought or other adverse weather phenomena will occur over the enemy territory thus depriving them of the necessary means of existence.

These wild ideas can never be realized, since atmospheric circulation takes place at the expenditure of an amount of energy which is incommensurable with the amount of energy at man's disposal.

At the same time, scientifically based experiments on influencing local meteorological processes are being carried out in the USSR, the USA, and other countries (V.Ya. Nikandrov, I.I. Gaivoronskii, G.F. Prikhot'ko, Langmuir, Schaeffer, and others).

Whereas the successful influence of atmospheric processes is for the time being still full of insurmountable difficulties, the modification of the climatic regime of the surface air layer is within man's reach.

Modification of the climate of the surface air layer

The climate of the surface air layer can to some extent be changed by changing the character of the underlying surface.

During his many ages of activity man has always been changing nature. In our country the changing of nature became a conscious, planned process. Huge projects are being carried out in Byelorussia, on the Volga, and in the Ukraine, in Transcaucasia and in Central Asia. In Belorussia fertile soils are being freed for valuable crops as a result of the draining of the huge Pinsk marshes. Over the North Caucasus and in Transcaucasia on the lower Volga, in the Ukraine, and in Central Asia irrigation channels are laid out for irrigating and bringing water to deserts and dry steppes. Field protecting forest belts are being created in steppe regions. All these measures help to raise the yield of agricultural crops, and to a considerable extent free agricultural production from weather hazards.

We have successfully limited the harmful affect of many natural phenomena, and transformed some of the destructive forces of nature into useful ones. In particular, advancing agrotechniques contribute to the improvement of the soil structure, which together with the field-protecting forest cultivation improves the hydrological regime, e.g., the runoff regime. Hydroelectric stations exploit considerable water resources and obtain electric energy for expanding navigation, irrigating large territories, and producing good harvests. The additional water arriving in the channels from humid regions, not only creates oases in deserts, but also moderates the heat and moistens the air. Field-protecting forest belts reduce the force of dry winds, harmful to agriculture, and modify the regime of the temperature, evaporation, and humidity of the air. By draining marshes the unproductive evaporation is reduced, resulting in dryer and healthier air.

How much will the climate of the surface air layer be modified as a result of all these measures? Although scientists are still doing research, it is already possible form a general picture of the expected climatic change.

Experimental investigations and theoretical calculations are being carried out to estimate this expected change. In the USSR these complex investigations are being performed by M.I. Budyko, O.A. Drozdov, D.L. Laikhtman, M.I. L'vovich, S.A. Sapozhnikov, M.I. Yudin, and others.

One of the main problems is: how does the melioration modify the regime of atmospheric precipitation?

Until results based on a rigorous scientific analysis were obtained various opinions were expressed in various popular pamphlets. Some considered that as a result of field-protecting forest belts, the amount of falling precipitation would sharply increase and dry steppes and deserts would be covered by a rich vegetation. Others denied any possibility of the forest belts modifying the amount of precipitation.

Investigations in recent years lead to the general conclusion that as a result of the melioration measures taken, only a very small increase in the atmospheric precipitation occurs, which cannot appreciably influence the moistening of the melioration regions. This is understandable, since the formation of clouds and precipitation is quite a complex process, depending on the character of the large-scale atmospheric circulation. With regard to the moisture content of the air, little evaporation over a bounded arid territory may have a comparatively small effect on increasing the moisture content of the air, since as a result of the constant motion of the air the moisture is continuously carried beyond the boundaries of this territory.

According to new data, the additional evaporation and intensification of the ascending currents over forest belts may increase the amount of precipitation over the southeast of the European territory of the USSR by on the average less than 15–20% of the annual amount of precipitation. Taking into account the fact that the mean annual amount of precipitation over the Lower Volga, to the east and south of Volgograd, does not exceed 300 mm, we conclude that the amount of precipitation there increases at best by 45–60 mm. However, if this increase does take place, this still does not mean that there will be a considerable increase in the humidity there. This may be due to the fact that the annual amount of precipitation oscillates within wide limits. In particular, in the region of Volgograd the annual amount of precipitation varies from 150–400 mm. Consequently, in a dry year, a small increase in the precipitation hardly reduces the dryness. The important thing in the struggle with drought is therefore not to "cause" a considerable amount of additional precipitation, but to modify the river runoff regime, to preserve the soil moisture, and to improve the climate of the surface air layer. This can be achieved by flooding, irrigation, cultivation of forest belts, and the application of specialized agrotechniques.

The supplying of water and irrigation of large arid soil areas provide considerable moistening of the fields. The arrival of water during the period of plant vegetation, particularly during dry years, considerably exceeds the amount of precipitation and thus secures the development of the plants and stable harvests.

Irrigation is not confined to moistening the soil and to creating optimum conditions for plant vegetation. The moisture which is absorbed by the soil evaporates. Since about 600 calories of heat are required for evaporating 1 gram of water, in the case of abundant irrigation a considerable part of the solar heat is used for evaporation, as a result of which the heating of the soil and of the adjacent air layers is noticeably reduced. Observational data obtained at meteorological stations situated in deserts and oases, as well as data obtained on the basis of experimental and theoretical investigations, show that the surface temperature of the soil of irrigated fields is 15°–20°C lower than that of nonirrigated fields.

The situation is somewhat more complicated with respect to the air temperature. The drop in the temperature over an irrigated section depends on its dimensions. If the irrigated section is a small oasis in the desert, it is natural that the air temperature there cannot be appreciably lower than the air temperature in the surrounding desert, since high-temperature air from the desert will arrive over the irrigated section. Owing to the small dimensions of the irrigated section, the air from the desert has no time to cool down. The results are different when the area of the irrigated fields is several square kilometers. In this case the air passing over the irrigated fields loses a part of its heat on evaporating the soil moisture and is appreciably cooled down. According to existing data, the mean difference in the air temperatures between the desert and such an oasis amounts to 2°–3°C. Over even larger areas of irrigated soil (tens and hundreds of square kilometers) the mean air temperature is 3°–5°C or more lower than over the desert.

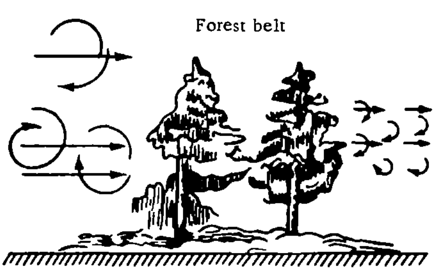

FIGURE 85. Influence of forest belts on the wind velocity and on the formation of air vortexes

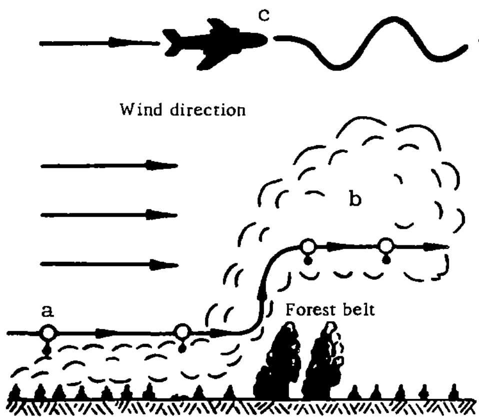

FIGURE 86. Vertical air motion over forest belts

The humidity over large irrigated sections also varies. Thus, the relative humidity of such sections rises by 20–30% as compared with the desert.

Forest cultivation in steppe and forest-steppe regions also contributes to the variation in the climate of the surface air layer. Forest belts obstruct the blowing of snow in ravines and gorges, and retain it on the fields thus giving additional moisture to the soil at the beginning of the vegetative season. They also somewhat weaken the force of snowstorms and dry winds. But this does not mean that dry winds are completely stopped, since they are generated by large circulation processes. Dry winds are not only observed in the surface layer, but extend over thick atmospheric layers, often up to a height of 10 km and more.

By planting forest belts on the edge of fields, the roughness of the ground surface is increased, thus slowing down the air motion (Figure 85). But the action of the belts is not limited to weakening the wind. Vortex (turbulent) motions constantly take place in the atmosphere. Small vortexes break down on encountering a forest belt and, passing through the belts, become even smaller. This is very important, since under higher turbulence the soil moisture evaporates more rapidly than under a low turbulence. Moreover, as is shown by experiments, air currents are created over forest belts, which, under an appropriate state of the atmosphere, may contribute to the development of cloudiness.

Figure 86 shows schematically the flight of a balanced rubber balloon (Figure 86 a) in a forest belt in the case of a wind directed perpendicularly to the forest belt. When it is sufficiently far from the forest belt the balloon flies horizontally to the ground surface. When it reaches the forest belt it moves upward drawn by the air current. Experience with smudging also shows that on reaching the forest belt the smoke ascends (Figure 86 b). The influence of forest belts is also felt by an aircraft flying at a height of 100–500 m (Figure 86 c). Both experimental data and theoretical calculations show that forest belts become more effective when there is a whole system of them at distances of up to several hundreds of meters.

Plants are already being prepared to make rivers entering the northern seas flow to the south in order to force them to give their water to the arid steppes of Kazakhstan. This too will bring about a variation in the climate of the surface air layer.

Recently a group of scientists headed by M.I. Budyko made a proposal to temper the climate of the northern countries by melting the ice of the northern Arctic Ocean. Calculations show that by reducing the reflective capacity of the arctic ice (albedo) by 10–20% it is possible to melt the entire ice cover of the ocean (the average thickness of the ice is 2.5 m). This can be achieved by covering the ice with a very thin layer of black powder.

During the summer months a larger total solar radiation arrives in the Arctic than in the equatorial zone. However, it is known that due to its high albedo the air temperature near the ice surface in the Arctic usually does not exceed 0°C. Over 70% of the solar radiation arriving there is reflected back into space, and the remaining part is lost on melting the ice surface.

The 30% of solar radiation absorbed is insufficient to completely annihilate the eternal ice of the ocean but if the albedo of this cover is reduced to 50–60%, the ice could be completely annihilated in the course of 2–4 years. It is true that during winter it will reappear, but by the end of May it will thaw again. In Central Arctic the air temperature near the water surface during summer will rise to 5°–8°C, and in certain places of the coastal zone – to 10°–12°C. As a result, on the extreme north of Asia, Europe, and North America the climate will become warmer than it is now.

It is not enough to aim only at increasing the air temperature in the Arctic and freeing the northern Arctic Ocean from ice, although navigation now takes place under heavy ice conditions. It is more important to determine whether the atmospheric circulation changes on a large scale, since if it does, the precipitation amounts change also, increasing in some places and decreasing in others, the latter bringing the weather phenomena harmful to agriculture.

CLIMATE FLUCTUATIONS

Extreme weather anomalies as one of the characteristics of climate

One often hears contradicting judgments on the variation of the climate. They are usually based on personal impressions. They are sometimes based on anomalously warm or cold winters or other seasons of the year repeating one after the other. Let us see if these judgments are well based.

It has already been said that the formation of the weather and of the climate are constantly affected by such factors as the inflow of solar energy, atmospheric circulation, and the underlying surface. However, in spite of these, during the same month or even season the winter is often different in different years. The characterization of a climate of a given place therefore involves both small and extreme deviations of the meteorological elements from the mean many-yearly values or norms. It follows that extreme deviations of the temperature or of other climate characteristics from the mean many-yearly values cannot serve as a basis for any conclusions on the variation of the climate.

Data of systematic meteorological observations, accumulated during the last 200 years show than an anomalous variation of the weather during a month or a season is quite an ordinary phenomenon. In particular, the winter mean monthly air temperature in countries with a continental climate, oscillates within wide limits. Thus, for example, in Moscow the winter mean temperature very often deviates from the norm by 2°–3°C, and in individual years, by even 6°–11°C. As a result the difference between the mean temperatures of anomalously warm and anomalously cold winter months may reach 20%. Compared to a mean many-yearly temperature for January of -10.3°C, the mean temperature of an extremely warm January may reach 2° to 3°C, and an extremely cold January, -20° to -22°C below zero. Thus, for example, in 1882 the mean temperature of January in Moscow was 3.0°C and in 1893 21.6°C below zero.

During the summer months in the central regions of the European territory of the USSR the deviations of the temperature from the norm do not exceed 3° to 6°C.

Similar temperature fluctuations take place also at other points of middle and high latitudes. Near the warm oceans these oscillations are considerably smaller, and at low latitudes they are negligible.

Sharp temperature fluctuations are associated with a deviation of other elements too from the norms, primarily the cloudiness and the amount of precipitation. It is known that in most cases a cold summer is characterized by more rain than a hot summer, a cold winter — by smaller amounts of precipitation than a warm one. It is true that there are also warm winters with a precipitation deficit. The latter is due to the fact that the transport of warm air from west or southwest gives rise to a positive temperature anomaly, but the prevailing anticyclonic circulation prevents the formation and fall of precipitation.

When the cyclonic activity is intense, a large amount of precipitation usually falls. Often the amount of precipitation during 24 hours equals the monthly norm and can exceed it. For example, in June, 1958 a large amount of precipitation fell in the central regions of the European territory of the USSR. On 27 June alone 58 mm of precipitation fell, which is almost the monthly norm, and during the whole month 122 mm fell, i.e., about double the monthly norm. The rainy month was associated with sharp temperature oscillations. In other years, on the other hand, a negligible amount of precipitation fell, equal to a few millimeters; sometimes no precipitation fell at all during an entire month.

A considerable weather anomaly is observed almost every year over some part of the terrestrial globe. Often even the mean monthly temperature deviates from the norm by 10° to 12°C over a large territory. Moreover, if a positive temperature anomaly is observed over one continent, over another a region of negative temperature anomaly is observed.

It has been established, for example, that between 1891 and 1934, characterized by considerable temperature anomalies, 20% of the anomalously cold and 18% of the anomalously warm months were simultaneously observed in Eurasia and America during January and February. In the remaining 62% of the cases anomalously cold or anomalously warm weather extended over only one of these continents during the month.

The same is true with respect to precipitation. This is particularly noticeable in the summer. Quite often there are cases where in western Europe cold weather is associated with the fall of abundant precipitation whereas at the same time in eastern Europe there is hot, dry weather. When there is cold and rainy weather in eastern Europe, in Siberia and western Europe, on the other hand, dry weather often prevails.

Observations also show that extreme values of the temperature of the same sign are sometimes repeated during a series of months. There are warm or cold winters, warm springs, cold or hot summers, and similar falls. Moreover, anomalously warm or anomalously cold winters, as well as other seasons, may be repeated successively over the same territory for 2-3 years. In these cases too, one cannot speak about climatic variations, since usually after an anomaly of a given sign an anomaly of the opposite sign begins.

Observed weather fluctuations are often identified with climate variations. Climate variations were "explained" by the reduction of the forest area or by the effect of the artillery during the first world war. In recent years the climate variation is "explained" by the explosion of atomic and hydrogen bombs, by the creation of water reservoirs and other measures for the transformation of nature. For example, a cold and rainy summer or a warm and snowy winter on the European territory of the USSR is "related" to the construction of such large water reservoirs as the Rybinsk, the Tsimlyansk, and others. On the other hand anomalously cold or hot summers, as well as anomalously cold or warm winters, were also observed earlier when the water reservoirs had not yet been constructed. There is no connection between the action of artillery or of explosions of hydrogen bombs and the climate variation. Large measures for the transformation of nature do lead to the modification of the climate of the surface air layer (to the variation of the temperature, humidity, wind, and others). These variations have a local character and cannot affect regions situated far from the place of the climate melioration.

Even if we consider the Volga Reservoir, whose area is over 300 km^2, it certainly exerts a noticeable influence on the coastal region moderating the summer heat and the winter cold. However, similarly to other small water basins, the influence of new water reservoirs shows up within several kilometers. The Moscow Sea, therefore, cannot exert any practical influence on the climate of Moscow. This is understandable if we take into consideration that, in addition to the underlying surface, such powerful factors as the inflow of solar radiation energy and the atmospheric circulation take part in the formation of the climate.

It still does not follow from the above that the climate does not change and that if it undergoes variations then these are only due to man's activity. There are many factors indicating the existence of natural variations of the climate both during geological periods, and during the course of the last centuries.

Organic and inorganic fossils give some idea about the climatic features of various geological periods. Thus, for example, coal deposits attest to the fact that the climate was humid during the cold period. Rock-salt deposits, on the other hand, indicate a dry climate. From rock floats it is possible to determine the motion of continental ice during the glaciation period. From the fossils of large trees it is possible to get an idea of whether the climate was characterized by a sharply displayed annual temperature march, was it cold or warm during the entire year, etc.

It is known that during the vegetative period annual rings are formed on trees. From the annual rings it is possible to get an idea not only about the character of the climate but also about the weather anomaly during individual seasons, since the thickness of the annual rings of trees depends on the degree of moistening of the soil. During dry years, due to the limited supply of moisture, the annual rings are thinner than those years with a sufficient amount of atmospheric precipitation.

Climatic variations of the geological periods

The last geological periods, the Tertiary and Quaternary, which lasted altogether 70 million years, are characterized by considerable variations in the climate. It was established that in the Tertiary period, equalling 69 millions of years, the climate was warm and humid over the present arid territory of the southeast of the European territory of the USSR, Evergreen heat-loving trees of the humid tropical climate used to grow there. In Europe it was also warmer than it is now. In the northern countries, broadleaf forests, characteristic of the temperate climatic belt, were widespread. It is assumed that the glaciation at high latitudes of the Earth began at the end of the Tertiary period and that it continued in the Quaternary period. The Quaternary period lasted for one million years.

The glaciation during the Quaternary period extended over land regions. In particular the territory of Europe up to 50° NL was covered by an ice layer up to 1 km thick. At this time in North America the ice was even thicker, and the southern glaciation boundary moved to 40° NL, and in certain places somewhat more to the south.

Later on the glaciers began to retreat to the north due to the rise in temperature. According to scientists, the retreat of the glaciers up to the south of Sweden occurred about 16,000 years ago. During the Quaternary period the ice either expanded to the south or retreated to the north as a result of a series of oscillations of the climate. Glaciation traces were observed also in the southern hemisphere.

On the basis of indirect data on the climate and vegetation, L.S. Berg showed that during the post-glacial period the successive warming up of the Earth, which was observed in the late-glacial period, stopped. The post-glacial period lasted for about 5000 years, i.e., between 13,000 and 8000 years ago.

Many different explanations are given for the climate oscillations. It is natural to assume that they are caused both by astronomical and by terrestrial causes. Some scientists assume that the climatic oscillations are related to the variation in the eccentricity of the Earth's orbit, to the displacement of the Earth's axis, to the activity of the Sun, etc. Others, that the climate oscillations are related to the variation of the outlines of the continents due to geotectonic processes, etc. L.S. Berg, has calculated that the temperature of the air and of the water during the glaciation period was 4° to 5°C lower than the present temperature even at the tropical latitudes.

However, the true causes of the climatic oscillations are still unknown.

Climatic variations of the last century

Whereas one has to guess at the climatic variations of the geological periods from indirect data, there are more accurate data about the climatic variations of the last 200 years. It was established by observations that during the last 100 years or more a gradual rise of temperature has taken place in the northern countries.

We have already said that the climate characterization involves any deviation of the meteorological elements from the norm. Figure 87 gives the curve of the mean monthly January air temperature for 155 years (1805–1960) from observations at Leningrad. It follows from this map that the mean monthly temperature undergoes sharp oscillations within quite wide limits. Thus, for example, in 1814 the mean January temperature was -21.4°C, whereas in 1866 it was only -1.4°C. It is difficult to observe the slow rise of temperature which took place during all this period from the temperature curve in Figure 87. To eliminate the sharp oscillations and find out the long-term march of the temperature, it is smoothed out by means of moving average quantities.

By means of a moving 35-year mean air temperature E.S. Rubinshtein showed that whereas in the first quarter of the 19th century the mean January temperature reached -10°C, in the first quarter of the 20th century it rose to -7°C; in other words, during the last century the air temperature in January rose by 3°C. The temperature rise was especially noticeable in the Arctic. In the 1920's the temperature of the water surface in the

Barents Sea rose by 8°C. In the northern Arctic Ocean the amount of ice decreased. Because of this, for the first time in history the icebreaker "Sibiryakov" passed the northern sea route from west to east in one navigation.

R. Scherhag, E.S. Rubinshtein, and others showed that the warming up in the 1920's and 1930's extended not only over the Arctic, but also over the adjacent regions of Europe, Asia, and North America.

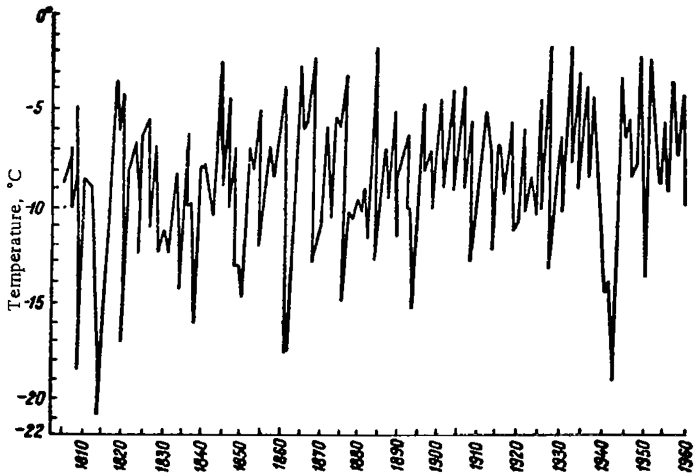

FIGURE 87. Monthly mean air temperature for January from 1805 to 1960 in Leningrad

Many investigators (M.S. Eigenson, L.A. Vitel's, G. Villet, and others) explain the climatic variations by oscillations of the solar activity, associated with variations of the atmospheric circulation. It has been established (V.Yu. Vise, L.A. Vitel's) that in the warming period in the north of Europe and in the western Arctic the cyclonic activity in the north of the Atlantic and of Europe was most intensive.

Variation in the level of the Caspian Sea and in the climate of the European territory of the USSR

A remarkable criterion for the climate variation of eastern Europe is the largest closed water reservoir in the world — the Caspian Sea. It is known that the level of the Caspian Sea continuously varies, these variations taking place more or less periodically. Figure 88 shows the long-term variation in the level of the Caspian Sea at the town Baku from 1556–1960. The graph was plotted by L.S. Berg on the basis of indirect data and direct observations which have been carried out since 1830. In recent years the graph has been complemented by us. The amplitude of the oscillation of this sea exceeds 5 m, and its level recently fell rapidly to the lowest level.

It can be assumed that the amount of water evaporating from the surface of the Caspian Sea is a more or less constant quantity. In this case it is

natural to assume that the variation in its level is connected with the an
of inflowing water. According to calculations of B.D. Zaikov, the wate
rich Volga brings to the Caspian Sea on the average 251.9 km^3 of wa
annually, which amounts to 77.7% of the total amount of water arriving.
The rivers Terek, Kura, and Ural bring 43.5 km^3 of water, or 13.4%,
all the remaining rivers taken together — only 28.8 km^3 of water, or 8.
The precipitation on the surface of the Caspian Sea amounts annually to 71.
of water. We can see that the Volga is the main source supplying the Caspia
Sea, since of the 400.8 km^3 of water arriving (including a small undergroun
flow) the Volga brings somewhat more than 60% of the total amount.

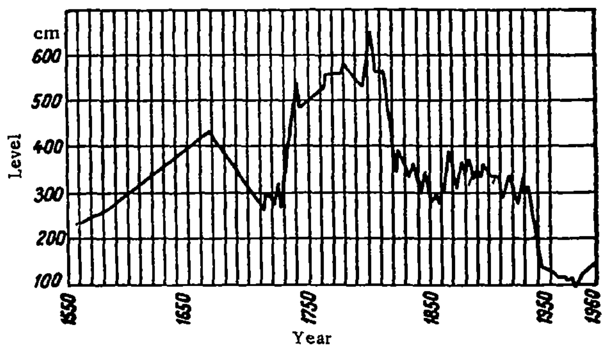

FIGURE 88. Long-term variation in the level of the Caspian Sea from 1556 to 1960

Many prominent scientists dealt with the explanation of the causes o
such large variations in the level of the Caspian Sea. Since the Caspia
is a closed basin, it is possible to determine its water balance, i.e., t
calculate the inflow and loss of water. The first attempt to determi
water balance of the Caspian Sea was made by A.I. Voeikov. N.M. Knipo
A.A. Kaminskii, B.D. Zaikov, and others subsequently determined the
balance more exactly. Kaminskii was the first to explain the variatio
the level of the Caspian Sea by the variation in the amount of winter pr
cipitation falling in the basin of the Volga. Subsequent investigations
firmed this conclusion.

Indeed, among the elements of the water balance of the Caspian Sea
important place is occupied by the runoff, first of all the runoff of the
More than half of the annual runoff of the Volga is connected with the s
high-water, which depends on the snow accumulation during the winter
months. Depending on the amount of winter precipitation, the runoff o
Volga oscillates within quite wide limits — from 300–350 km^3 during y
abundant in precipitation to 150–200 km^3 during years with little snow.
large oscillations of the Volga runoff must influence the level of the Caspi

The runoff of the Volga decreased considerably recently due to a s
of winters poor in snow. According to approximate calculations, durin
period 1930–1945, owing to the precipitation deficit, the runoff of the Vol
creased by 590 km^3, which includes a decrease of 551 km^3 due to the preci
tion of the winter period. If we take into account that the annual evaporati
from the Caspian Sea amounts to about 400 km^3, then this decrease in the
cipitation and in the runoff should lead to an appreciable lowering of the le
This lowering of the level has been observed during the last 20–25 year

The question arises: what will happen to the Caspian Sea in the future? Will the sea level continue to fall? L.S. Berg, studying the character of the oscillations of the Caspian level, arrived at the conclusion that the fall of its level is related to the warming up of the Arctic, caused by intensification of the cyclonic activity to the north and of the anticyclonic activity over the European territory of the USSR. Consequently, the oscillations of the level of the Caspian Sea are determined not by local weather conditions, but by large-scale atmospheric processes, causing the climate oscillations. According to the data of B.D. Zaikov, a small amount of precipitation and a decrease in the runoff during the period 1930–1945 was observed simultaneously in the basins of the Don, Ural, Irtysh, and Tobol. At the same time, the mean runoff was particularly high in the basins of the Dnieper, Angara, and Shilka.

As to the level of the Caspian Sea, since its variations are connected with the climatic variations and have a more or less periodic nature, a new rise in its level can be expected in the future. However, it must be taken into consideration that due to the melioration of the climate of the arid southeast of the European territory of the USSR a part of the Volga water will be spent on irrigation and supplying water to large soil areas. This will lead to a decrease in the runoff, which no doubt will be reflected in the general fall of the level of the Caspian Sea. It should be said that works on irrigation and water supply of fields are carried out within the boundaries of the basin not only of the Volga, but also of the Kura, Araks, Terek, and other rivers.

Other ideas have recently been proposed about the reasons for the oscillations in the Caspian level. In particular, it is assumed that the variation in the level of the Caspian Sea is also connected with geological processes — the lowering and rising of the sea bed.

Studying the climate of the Russian Plain during the historical period, I.E. Buchinskii arrived at the conclusion that the climate there undergoes long-term oscillations. He succeeded in establishing that harmful natural phenomena were most often observed over this territory during the 15th century, when, as a result of the climate disturbance, the number of dry years, of severe winters, of storms, etc., increased. However, although it underwent some fluctuations during the last 2000 years or more the climate has undergone appreciable changes. Thus, in the time of Herodotus, who lived from 484 to 425 B.C., the climate of the Russian Plain was in general outlines the same as it is today.

Long-term climatic oscillations, independent of man's activity, take place owing to the variation in the character of the general circulation of the atmosphere. In turn, the character of the circulation obviously depends on the solar activity and on other astronomical causes. But the mutual relation and the mechanism of their action on the circulation of the atmosphere still remain one of nature's big secrets. This problem is not only of cognitive interest. The physical reasons for the variation in the general circulation of the atmosphere is the most important scientific and practical problem in meteorology, since only when the laws of the atmospheric circulation are known, will it be possible to develop hydrodynamic methods of short- and long-term weather forecasting.

The study of the physical processes taking place in the atmosphere and the determination of the laws of the general atmospheric circulation make it possible not only to work out methods of weather forecasting, but also in the long run to control the weather and modify the climate.